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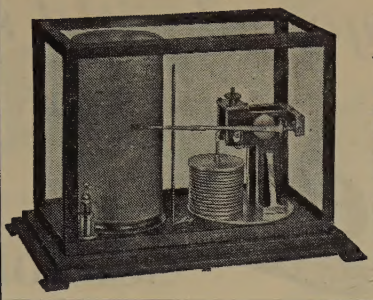

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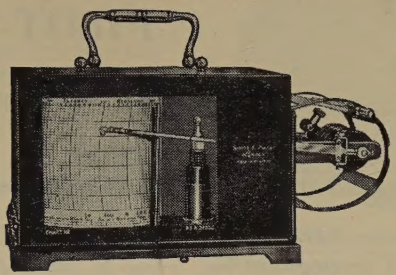

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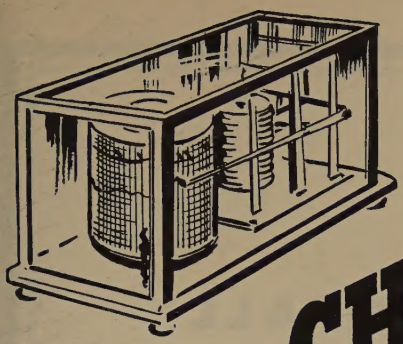

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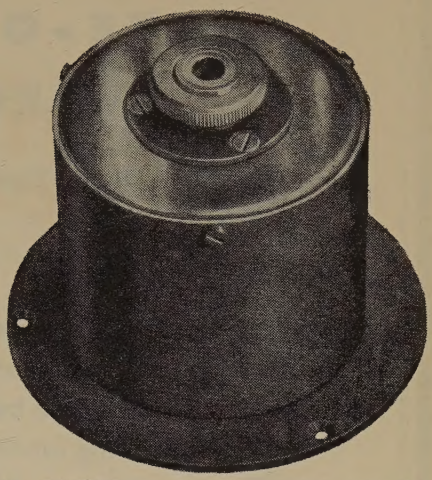

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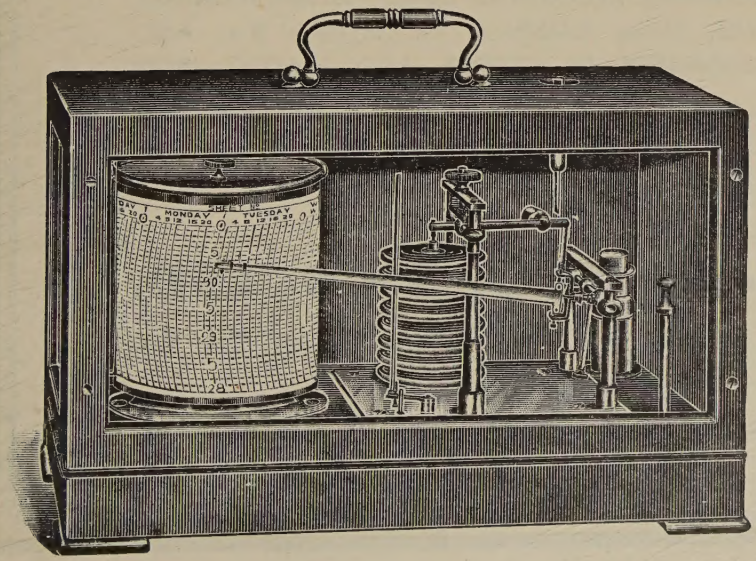

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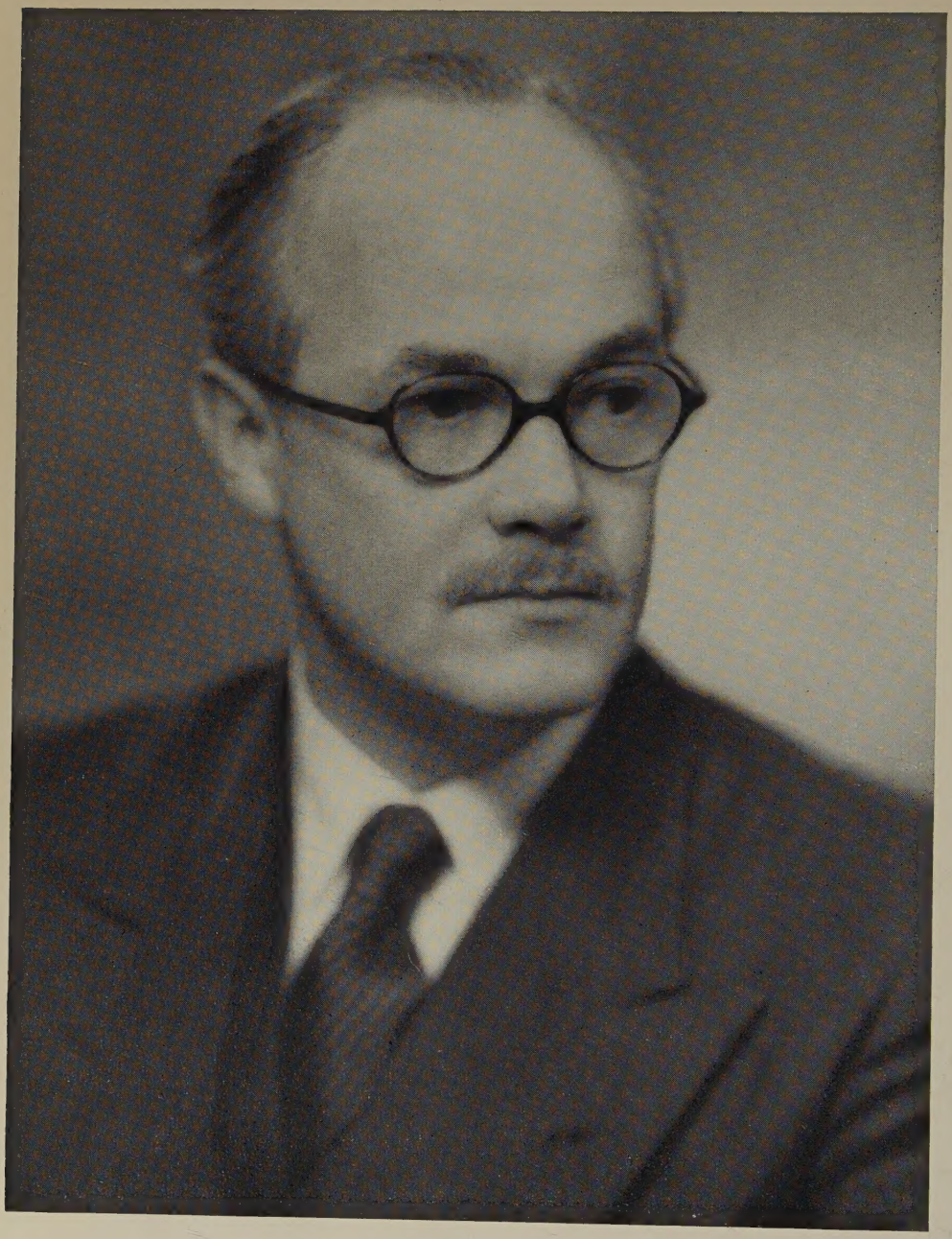

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Winds and temperatures between 20 km and 100 km – a review

By R. J. MURGATROYD
Meteorological Research Flight, Farnborough, Hants.

(Manuscript received 28 May 1957)

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1. Introduction

During the last two decades measurements of temperature and wind in the troposphere and lower stratosphere by radiosonde and radar wind-finding methods have become commonplace in many countries and the main features of the temperature distribution and zonal wind circulation over the world in these regions are now quite well known (Kochanski 1955, Bannon 1954, 1955). The amount of data available for greater heights is much smaller and, up to the present, few measurements from balloon-borne instruments are available for heights above 30 km and none above about 40 km. A few other direct observations have, however, been obtained from high-altitude rocket experiments and a

good deal of information on winds and temperatures can be deduced indirectly from other experiments based on phenomena such as sound and radio propagation through the upper atmosphere and observations of the solar spectrum. In addition certain naturally occurring phenomena such as meteors, the aurora and noctilucent clouds have provided valuable information, so that, with the help of theoretical considerations, it is now possible to build up a general picture of winds and temperatures over the globe up to considerably higher levels. Various attempts have been made to do this up to 80-100 km during the last few years (e.g., Kellogg and Schilling 1951, Chiu 1951, Haurwitz 1954, Pant 1956). At least one attempt has also been made to extend the picture to greater heights (Deb 1953). It is the purpose of this review to outline and then synthesize all the information, published before 1957 and known to the author, into a reasonably consistent system of zonal winds and temperatures up to an altitude of 100 km. Insufficient information is available as yet to do this for greater heights and such as exists will only be mentioned briefly. It is hoped that this will provide a general background to which the measurements of the International Geophysical Year can be added.

2. Theoretical considerations

2.1. Temperature distribution and its control by the radiation field

The three-dimensional distribution of temperature and air motion depends on the energy balance. Neglecting friction, and assuming a steady state :

Energy from solar radiation + energy from terrestrial radiation + energy from other sources + energy due to motion (advection + convection) = 0.

In this section these four energy terms will be briefly considered.

The first term is the most fundamental and, fortunately, owing to recent improvements in our knowledge of the solar ultra-violet spectrum, can be estimated with reasonable confidence. The second term is more difficult to evaluate but methods and data are rapidly improving. Throughout most of our region of interest the third term is small (unlike the troposphere where there is a large latent heat term). The energy due to air motion is indeterminate at present and our knowledge of it depends on an evaluation of the other terms. It may often be that the last two terms are small compared with the first two and in this case the atmospheric temperature will not greatly differ from the radiative equilibrium temperature. Gowan (1947) has made a series of calculations based on this assumption and obtained temperature structures which agree reasonably well with those found by measurement. It should in principle be possible to compute atmospheric motions and their subsequent thermal effects from an initial state of radiation equilibrium found in this way. Goody (1954, p. 167) and Sheppard (1954) considered that, although it is not likely that radiative equilibrium exists anywhere at the levels under consideration, the sign of the lapse rate of temperature and a first approximation to the temperature itself may well be derivable from this assumption. It is probable also that the sign of derived horizontal temperature gradients would generally also be correct.

Solar radiation of different wavelengths penetrates varying depths of the atmosphere and suffers maximum absorption at different heights (Johnson, Purcell, Tousey and Watanabe 1952). Wavelengths below about 1800 Å are absorbed above 100 km apart from various windows between 1,000 Å and 1,300 Å and in the X-ray region where there is penetration to 80 km or below (Ditchburn 1956). Wavelengths between 1,800 Å and 2,900 Å reach about 35 km and longer wavelengths reach the earth's surface (Tousey, Watanabe and Purcell 1951). Molecular oxygen absorbs strongly around 100 km in the wavelength range 1,300-2,000 Å. There is little absorption of solar radiation around

80 km. Below about 70 km absorption by ozone becomes increasingly important, particularly between 2,000 and 2,900 Å. The absorption coefficients are so large that most of the solar energy is absorbed high in the tail of the ozone distribution so that in the main ozone layer (15 to 35 km) only weak bands in the near ultra-violet and visible can absorb energy. Absorption of infra-red solar energy by water vapour is important at lower levels (Karandikar 1946).

Since the solar intensity in the different wavelengths is known from rocket experiments and the amounts of molecular oxygen and ozone can also be determined approximately from the results and from calculations of photochemical equilibrium, the rate of heating due to absorption of solar energy at various levels between about 100 and 30 km can be computed (Craig 1951, Johnson 1953). A computation of this type (Murgatroyd 1956b) is outlined. The appropriate expressions are (Goody 1954) :

$$\Phi_\lambda \, \Delta\lambda = S_\lambda \, \Delta\lambda \cos\psi \exp\left(-\int_z^\infty k_\lambda \rho_r \sec\psi \, dz\right) \quad . \quad . \quad . \quad . \quad (1)$$

and

$$\frac{dQ}{dt} = c_p \rho \frac{dT}{dt} = -\int \frac{d\Phi_\lambda}{dz} d\lambda \quad . \quad . \quad . \quad . \quad . \quad . \quad (2)$$

where $S_\lambda \, \Delta\lambda$ = flux of solar radiation perpendicular to the beam outside the atmosphere in the wavelength range λ to $\lambda + \Delta\lambda$

$\Phi_\lambda \, \Delta\lambda$ = vertical flux of solar radiation at altitude z in this wavelength range

k_λ = absorption coefficient

ψ = solar zenith angle

ρ_r = density of absorbing gas

$\frac{dQ}{dt}$ = rate of heating per unit volume

ρ = atmospheric density

c_p = specific heat of air at constant pressure

$\frac{dT}{dt}$ = rate of change of temperature

TABLE 1. DENSITIES OF AIR, OXYGEN AND OZONE (MOLECULES/CM^3) USED IN CALCULATING FIG. 1.

Height km	Air density (molecules/cm^3)		Oxygen density (O_2 molecules/cm^3)		Ozone density (O_3 molecules/cm^3)	
100	2·3	10^{13}	6·3	10^{11}	–	–
95	4·5	10^{13}	6·3	10^{12}	–	–
90	8·7	10^{13}	1·8	10^{13}	3·4	10^{6}
85	1·8	10^{14}	3·8	10^{13}	1·3	10^{7}
80	3·9	10^{14}	8·2	10^{13}	4·6	10^{7}
75	9·9	10^{14}	2·1	10^{14}	1·7	10^{8}
70	2·5	10^{15}	5·3	10^{14}	5·9	10^{8}
65	4·5	10^{15}	9·5	10^{14}	1·9	10^{9}
60	8·2	10^{15}	1·7	10^{15}	5·6	10^{9}
55	1·4	10^{16}	2·9	10^{15}	1·8	10^{10}
50	2·3	10^{16}	4·8	10^{15}	5·1	10^{10}
45	3·9	10^{16}	8·2	10^{15}	1·4	10^{11}
40	6·8	10^{16}	1·4	10^{16}	4·4	10^{11}
35	1·6	10^{17}	3·2	10^{16}	1·1	10^{12}
30	3·9	10^{17}	7·8	10^{16}	3·6	10^{12}

Data for the solar intensities and absorption coefficients have been listed by Craig (1950) and Table 1 gives representative values of ozone and oxygen densities at the various altitudes based on the report of the Rocket Panel (1952) and theoretical calculations of Bates and Witherspoon (1952). According to Craig (1950) the ozonosphere above 35-40 km is substantially in photo-chemical equilibrium and the ozone amount above 35 km not very dependent on solar zenith angles. If this is so the data are substantially correct for all latitudes.

The results given in Fig. 1 demonstrate that maxima of heating due to absorption of solar radiation occur between 40 and 50 km and around 100 km with a minimum around 80 km. The heating decreases, and the altitudes of the maxima increase as the solar zenith angle increases. Fig. 2 gives the latitudinal distribution from summer to winter pole, illustrating maximum values at high latitudes in summer decreasing through smaller values at low latitudes to no heating during the polar night. The altitude of maximum heating is lowest at low latitudes. The oxygen distribution given in Table 1 is less certain than that of the ozone and the results of the calculations around 100 km should be treated with reserve. Pressman (1955) has made somewhat similar computations for the ozonosphere with different ozone amounts and concluded that the latitudinal

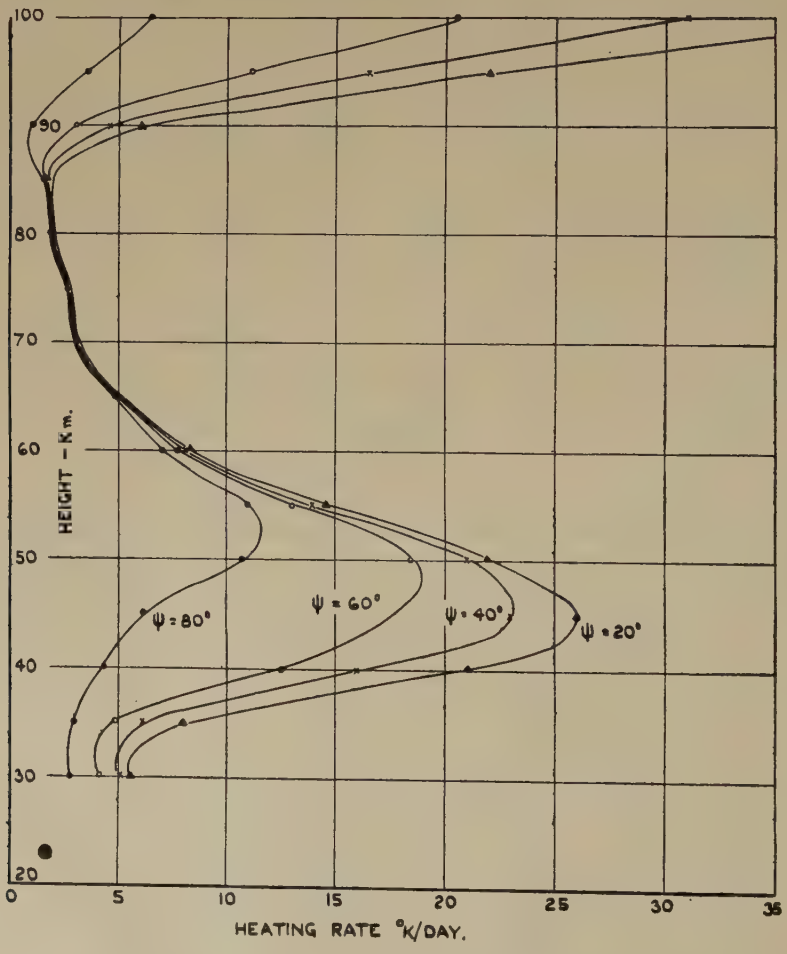

Figure 1. Variation of solar heating with zenith angle ψ.

heating pattern is not very dependent on the total amount of ozone present (although, of course, its vertical distribution is important).

Calculations of atmospheric and terrestrial radiation are more complicated than those only involving solar radiation. Carbon dioxide, ozone and water vapour are the important constituents and the complexity of their absorption spectra, the variation of their absorption coefficients with temperature and pressure, and their different distributions in the atmosphere lead to extremely lengthy computations (Möller 1951, Goody and Robinson 1951). Although the amounts of ozone and carbon dioxide (which probably has a constant mixing ratio of 0·4 — 0·5 g/kg) can be estimated at all levels of interest here, there are very few data on the water-vapour content of the ozonosphere. The aircraft measurements of Murgatroyd, Goldsmith and Hollings (1955) to about 16 km and a few balloon measurements of Barrett, Herndon and Carter (1949) to 30 km indicate that the mixing ratio is between about 0·002 and 0·05 g/kg in the lower stratosphere. If these comparatively low values persist into the ozonosphere the effect of the water vapour there will be relatively small. Some success has attended attempts to explain the main thermal features of the lower stratosphere by studies of the infra-red radiation balance there (e.g., Dobson, Brewer and Cwilong 1946, Goody 1949). For higher altitudes in the ozonosphere Plass (1956a, 1956b) has made specimen calculations of radiative transfer for ozone and carbon dioxide with given temperature structures and amounts of the constituents, demonstrating that the infra-red radiative transfer results in cooling approximately sufficient to offset the heating from the absorption of solar radiation. Curtis (1956) has proposed a simplified method of making this type of calculation and Curtis and Goody (1956) have produced specimen results for carbon dioxide based on it and also have shown how these computations can be extended to altitudes above 80 km or so where Kirchoff's law is no longer valid. Before any attempt can be made to investigate the distribution of cooling rates with latitude it is necessary also to know the mean temperatures at the different altitudes in each latitude.

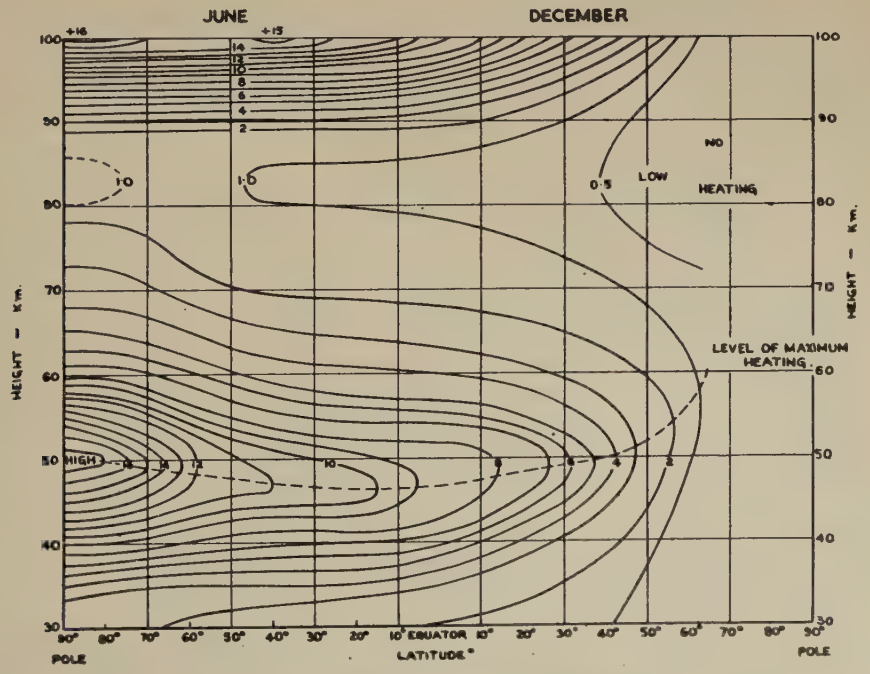

Figure 2. Heating due to solar radiation °K/day.

Summarizing, the results up to date of the theoretical studies of the radiation field suggest that :

(i) The temperature will rise to a maximum in the ozonosphere at about 50 km. Above that level it will decrease with height to a minimum around 80 km and then increase with height up to at least 100 km.

(ii) The temperature will show a large seasonal change at high latitudes and little variation in low latitudes, the horizontal gradient having the same direction continuously from summer pole to winter pole. This gradient will result in an increase of westerly wind components with height in winter and the reverse in summer.

(iii) The accuracy of the calculations is not yet sufficient to estimate the magnitude of the difference between the heating and cooling rates or even its sign, except perhaps at high latitudes where seasonal changes are large. During the polar night, when there is no solar heating, advection together probably with subsidence (Kellogg and Schilling 1951) will be important near the top of the ozonosphere. Depending on the lapse rate there may be considerable vertical heat transfer between about 50 and 80 km.

(iv) Wind motions may also have some feedback effect on the heating and cooling patterns by redistributing the radiatively important constituents (Sheppard 1954).

2.2. Other physical factors affecting the wind field

An exhaustive list of factors likely to affect the wind field at all altitudes has been given by Chapman (1954). Those which may be of major importance between 20 and 100 km are briefly considered in this section :

(i) The equations of motion will contain the pressure gradient, geostrophic and cyclostrophic forces in the same way as at lower levels.

(ii) Friction effects are negligible in the lower stratosphere but become more important at higher altitudes. The molecular kinematic viscosity at 50 km is about 2×10^2 cm^2 sec^{-1} increasing to about 1×10^5 cm^2 sec^{-1} at 100 km. The latter value is of the order of that of the eddy kinematic viscosity in the lower troposphere.

(iii) It is possible that surface topography can produce wave motions up to very considerable heights and there is some evidence that mother-of-pearl clouds up to 25 km may be produced in this way. Such effects, however, would be limited by atmospheric conditions at lower altitudes.

(iv) Any effects arising from the latent heat of water are probably very small. Although mother-of-pearl clouds are sometimes reported in high latitudes (often associated with very low temperatures in the stratosphere) the measurements up to date in mid-latitudes have indicated low humidities up to 30 km. No other clouds have been observed in the ozonosphere although noctilucent clouds at about 80 km are sometimes seen in high latitudes. There is considerable doubt, however, whether these are ice clouds or dust (Vestine 1934, Ludlam 1955). Dust arising from meteors entering the earth's atmosphere may accumulate near the base of the inversion at 80 km and can be observed under suitable conditions of illumination (Bigg 1956). This might have some influence on thermal conditions due to its absorption of solar radiation.

(v) Chapman also pointed out that the energies of dissociation, ionization and excitation, which are forms of latent heat, may have considerable importance in the dynamics of the ionosphere. These processes, however, are limited to altitudes above about 80 km.

(vi) Around and above 100 km magneto-hydrodynamic forces have to be considered. Their existence has long been established from the regular diurnal ' quiet-day ' variations in surface observations of the earth's magnetic field. The dynamo theory originally

proposed by Balfour Stewart (1882) to explain these variations requires horizontal movement by winds of an atmospheric conducting layer across the magnetic field of the earth. Solar and lunar diurnal variations are observed and are probably due to tidal and thermal effects. Vestine (1954) using the dynamo theory has attempted to deduce seasonal winds in the upper atmosphere and has obtained some agreement with the circulation system proposed by Kellogg and Schilling (1950). Other theories, e.g., Ross Gunn (1928), Chapman (1929), however, have been advanced which do not require wind motion to explain the observed magnetic variations and the dynamo theory, although at present widely accepted, has not been completely confirmed.

(vii). The effect of atmospheric oscillations or tides can be detected in records of surface pressure. A harmonic analysis of the daily pressure variation at various places over the earth in terms of a Fourier series shows that there are four main terms corresponding to 24, 12, 8 and 6 hour periods. The diurnal wave, which is mainly due to solar heating, is a progressive wave travelling from east to west round the earth with the sun. The semi-diurnal wave, which is probably largely due to gravitational effects, although thermal effects are also important, comprises two components and may be expressed as the sum of a progressive wave travelling round the world in 12 hr plus a standing wave (Simpson 1918). The former component is the larger and is most important near the equator. The latter is important near the poles and produces a pressure variation which at a given instant is the same everywhere on a parallel of latitude. The 8-hourly oscillation is comparatively minor and the 6-hourly oscillation only just traceable. To each pressure variation there is a corresponding air motion, and periodic winds accompanying the pressure oscillations can be computed (e.g. Stolov 1955).

It might have been expected that the lunar semi-diurnal tidal component would be easier to detect than that due to the sun, as the moon's gravitational effect is 2·2 times that of the sun. Instead, the observed amplitude ratio is about 0·06 and the lunar component was difficult to find (Chapman 1918). In order to explain this, the resonance theory first suggested by Kelvin has been developed. It was known from observations of the speed of propagation of the Krakatoa wave (Symons 1888) that the atmosphere has a free period of 10·5 hr, but Taylor (1936) showed that the atmosphere has more than one free mode of oscillation. Pekeris (1937), assuming a vertical temperature distribution in which, above the tropopause temperature minimum and isothermal stratosphere, there is a temperature maximum at about 60 km with a lapse above to a minimum at 80 km (thus postulating a further barrier capable of trapping energy passing through the stratosphere) was able to calculate that the two modes of oscillation were about 10·5 hr and 12 hr respectively. A natural resonance with 100-fold magnification would occur with a free period within about 6 min of the 12 hr. The results of Pekeris show a node at about 30 km. Above and below this level the atmospheric oscillations are in the opposite phase, and the phase at 100 km agrees with that required by the dynamo theory (southerly components at about 0700 hr). The specific kinetic energy ($\frac{1}{2}\rho V^2$) does not vary much with altitude so that the velocity of the semi-diurnal wind component increases from a few cm/sec near the surface to tens of m/sec near 100 km. Similar results have been given by Weekes and Wilkes (1947). Further information on this subject may be obtained from the reviews by Wilkes (1949, 1952) and Chapman (1951).

It may be concluded from the above considerations that it is valid to apply the equations of motion used in the troposphere up to 70 km or so. At greater heights tidal effects must be included and magneto-hydrodynamic forces, energy transfers involved in dissociation, ionization and excitation, and large increases in the molecular kinematic viscosity assume increasing importance and cannot be neglected above 100 km. The interpretation of observations of air motion at very high levels must take account of these factors.

3. A SUMMARY OF EXPERIMENTAL METHODS AND AVAILABLE MEASUREMENTS

3.1. *Introduction*

Many of the observations described below lead to values of atmospheric pressure p, density ρ, scale height H, or speed of sound a, and the temperature T must be derived from them. This is usually done by using the hydrostatic equation, the equation of state and the equation for the speed of sound in a perfect gas which are listed below

Scale height

$$H = \frac{RT}{mg} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

where R is the universal gas constant, g is the gravitational acceleration and m is the mean molecular weight. H is the equivalent height of a homogeneous isothermal atmosphere of temperature T.

The hydrostatic equation can be written in the form

$$\frac{dp}{p} = -\frac{dz}{H} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

where z is the height.

The equation of state is

$$p = \rho \frac{RT}{m} = \rho g H \quad . \quad . \quad . \quad . \quad (5)$$

The speed of sound for an infinitesimally small disturbance in a continuous medium is

$$a = \sqrt{\gamma \frac{RT}{m}} = \sqrt{\gamma g H} \quad . \quad . \quad . \quad . \quad (6)$$

where γ is the ratio of specific heats. In all these formulae it is not possible to separate T and m and hence the value of the temperature found always has to depend on the assumed value of m. Usually m is taken as 28·97, the surface value for dry air, to at least 70 or 80 km, the composition of the atmosphere being virtually constant up to this altitude (Chackett, Paneth and Wilson 1950). At greater altitudes lower values of m should be used owing to the dissociation of molecular into atomic oxygen (Rocket Panel 1952).

In the literature describing experimental determinations of wind there are two conventions as regards wind direction. Ionospheric workers usually define it in terms of northwards, southwards, etc., i.e., as direction away from an observer. Meteorologists regard northerly, southerly, etc., as motion towards the observer. In this paper the meteorological convention is used throughout and a north wind or south wind is regarded as a wind coming from the north or south respectively towards the observer.

3.2. *Balloon measurements*

During the last few years balloons have been developed capable of carrying radiosonde equipment and radar wind-finding targets to 25-30 km for routine measurements of temperature, pressure and wind. Larger balloons have also been developed for carrying to these altitudes heavier equipment for studies of atmospheric composition, solar radiation electrical conductivity and cosmic rays. The highest reported measurements of temperature and wind using balloons extend to about 42 km (Conover and Wentzien 1955).

Temperatures are usually measured directly by means of resistance thermometers, thermistors or bimetallic strips whose indications are telemetered to the ground. Errors are probably only a few degrees K mainly owing to radiational effects. Theoretically the temperature T could also be deduced from the measurements of height and pressure

but this at present is not as accurate as the direct measurement. An ascent to 30 km may take about 2 hr during which the balloon may drift considerable distances. In interpreting any series of results care has to be taken that they are not biased towards occasions of comparatively light winds, as there is often the possibility that the balloon may drift beyond the range of the telemetering link when the winds are strong.

Using a few years' reported results, several authors have constructed mean hemispherical cross-sections of temperature during different seasons along given longitudes. Good examples are those of Kochanski (1955) for 80°W up to about 30 km. By means of the hydrostatic and geostrophic wind equations, corresponding zonal wind fields have also been constructed. These data are large-scale means and take no account of meridional components, diurnal changes, longitudinal variation and smaller-scale systems, all of which have been noted in the literature (e.g. Bannon and Gilchrist 1956, Kochanski and Wasko 1956, Kochanski 1956, Fletcher 1954). The principle features above about 20 km illustrated by these cross-sections are outlined below. Detailed values are included in Figs. 3 and 4.

(i) In summer, there is an inversion between 20 km and 30 km at all latitudes, and the temperature in high latitudes is greater than that in low latitudes. The horizontal temperature gradient is greatest in middle or high latitudes. There is a gradual wind change with height from the tropospheric westerlies to an entirely easterly flow at 30 km.

(ii) In winter, conditions in low latitudes are similar to those of summer. Near the poles, however, temperatures are much lower, the annual range at 30 km, being 40°-50° K. Between 20 and 30 km there is an inversion in low latitudes, with roughly isothermal conditions in mid-latitudes and a lapse of temperature in high latitudes. The horizontal gradient of temperature varies considerably with latitude, the mid-latitudes being warmer than either the poles or the equator at 20 km : towards 30 km it is small in low latitudes but large in high latitudes, with the lowest temperatures over the poles. The effect of this temperature distribution is to reduce the tropospheric westerlies to considerably lighter winds in the lower stratosphere and then for the westerly components to increase with altitude towards 30 km, the increase being most marked in high latitudes.

(iii) The overall picture of variation of zonal wind components near 30 km in mid and high latitudes is therefore a monsoonal type of change from westerly winds in winter to easterly winds in summer, the former being the stronger. This is consistent with the existence of a circumpolar vortex, cyclonic in winter and anticyclonic in summer. In the mid-season months, conditions are intermediate between those of summer and winter, the pattern of the wind and temperature fields in April perhaps tending to be more similar to the summer pattern and that of October more similar to the winter pattern.

In low latitudes it is not possible to use the geostrophic equation to relate the temperature and zonal wind fields. Conditions over the equatorial regions with particular reference to the Pacific area, have been described by Palmer (1954). Between about 20 and 25 km there is a narrow band of westerlies – the ' von Berson westerlies ' – but at higher altitudes these are replaced by a broad easterly stream with a speed of about 30 m sec^{-1}. This was first observed in 1883 from the movement of the dust thrown up by the Krakatoa explosion (Symons 1888) and later confirmed by balloon ascents. This easterly airstream at about 30 km almost certainly extends to considerably higher latitudes in the summer but it has not yet been established whether the easterly winds at this altitude in winter at 80°W shown in Fig. 4 are found in all longitudes.

Mean values of temperature and their standard deviations up to about 30 km have been listed by McNaughtan (1956) using data produced by the British Meteorological Office. The mean values are similar to those in Fig. 3. The standard deviations increase with height and latitude in winter : in summer they are considerably less and show little

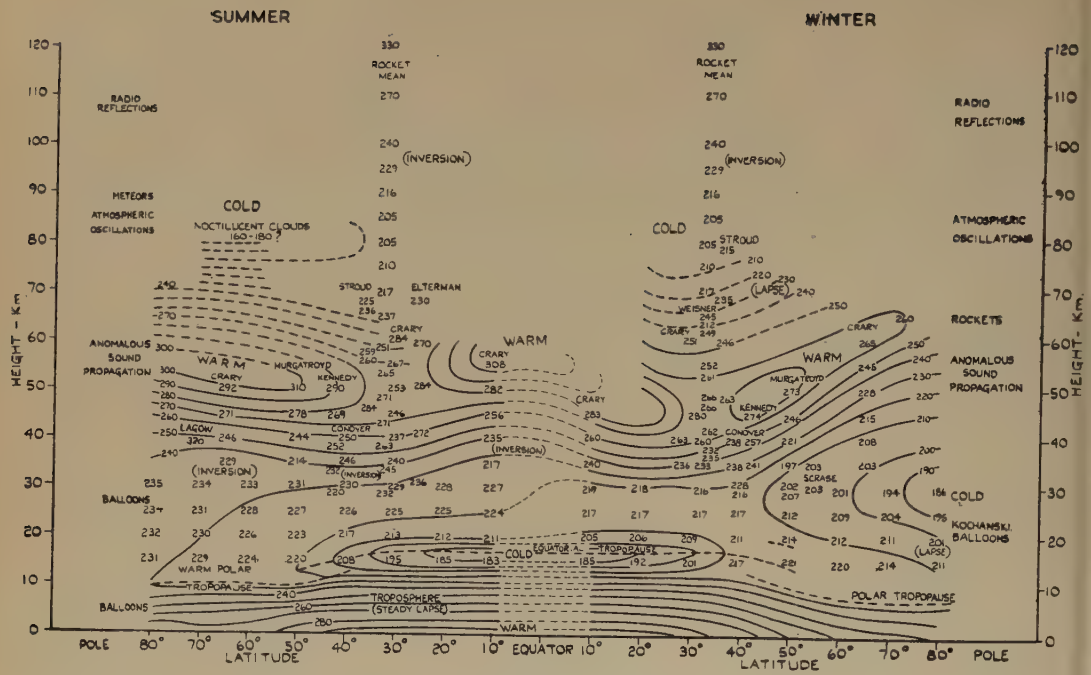

Figure 3. Observed temperatures °K.

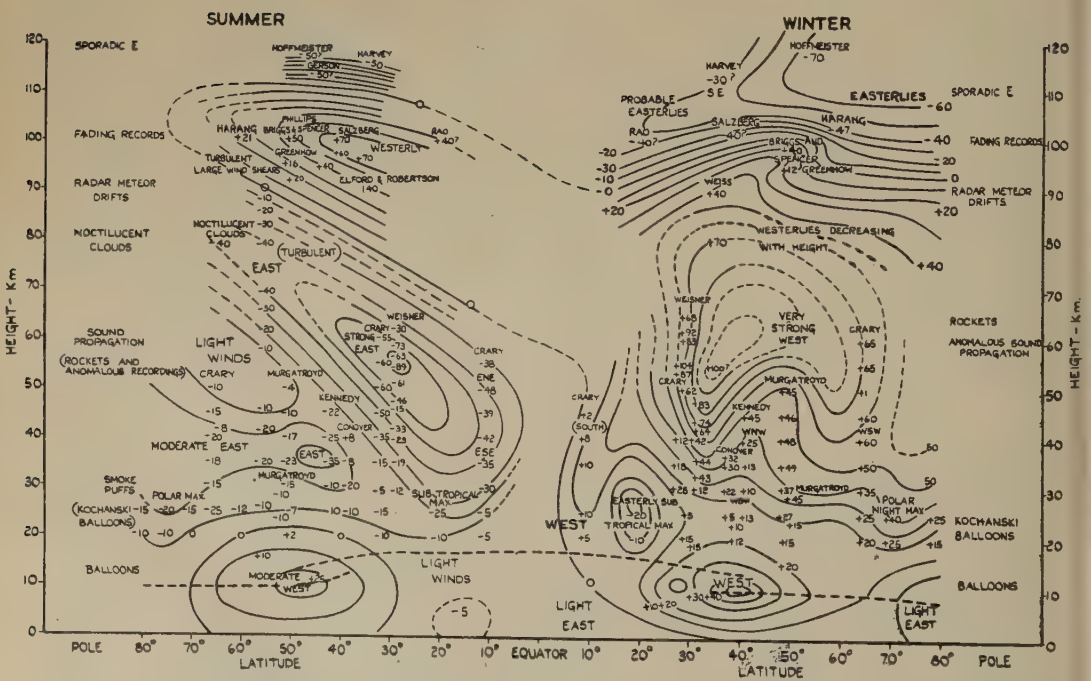

Figure 4. Observed zonal wind component. Westerly components positive m/sec.

variation. Typical values are given in Table 2. Kochanski (1956) gives 2°C in temperature and about 6 m sec^{-1} in vector wind as the average changes in 24 hr in mid latitudes between 15 and 30 km.

TABLE 2. STANDARD DEVIATION OF TEMPERATURE AT DIFFERENT ALTITUDES AND LATITUDES (°K)

	Winter				Summer		
Latitude / Height (km)	70°	50°	30°	0	30°	50°	70°
31	13	11	6	6	6	4	4
24	12	10	6	5	5	4	4
18	8	7	5	5	5	3	3
15	6	6	4	3	4	3	3

Results similar to those obtained from these general studies have emerged from individual sets of measurements. Scrase (1954), from a study of 150 soundings at Downham Market, England, and Lerwick, Scotland, found the annual range of temperature to increase with height to about 35°K. Scherhag (1952), found annual ranges at Berlin of 15°C at 20 km rising to more than 40°C at 40 km. Scrase, Scherhag, Kochanski and Palmer have also noted independently that large variations from the mean temperature and winds at these levels can occur for periods of several days. Brasefield (1954) has also reported variations of about 20°K at 30 km during the period of a month. Other notable individual studies of high-level balloon ascents are those of Ockenden (1939), Hulburt (1947), Gutenburg (1949), Brasefield (1950) and Conover and Wentzien (1956). The results of all these fit well into the general picture. It seems that the broad outline of the wind and temperature structure up to 30 km is now clear : the details should be filled in within the next few years.

3.3. Drift of smoke from shells and rockets

Murgatroyd and Clews (1949) made a series of measurements of winds at 30 km in England by observing with theodolites the drift of smoke produced by shell bursts. This method, besides being expensive, is restricted to clear weather. They showed that the wind was mainly westerly at 40-50 m sec^{-1} in the winter and easterly about 15 m sec^{-1} in summer. More recently in the U.S.A. smoke emitted from high-altitude rockets has been used by Kellogg and Schilling (1950) who reported a westerly wind of 90 m sec^{-1} at 35 km in December 1948. Further improvement of this technique can be obtained by emitting radar reflecting particles and tracking their drift, avoiding the restriction of clear weather required for visual tracking. The emission of a cloud of sodium vapour from a rocket in connection with twilight and night air glow studies has also provided a method of wind measurement. Over New Mexico in Oct. 1955 one was observed to drift in a north-west wind of about 90 m sec^{-1} at 85 km while its drift at 110 km indicated a wind there of about 50 m sec^{-1} from the south-east (Armstrong and Dalgarno 1955, p. 122). A similar experiment in January 1955 indicated strong west winds in the 50 to 70 km region.

3.4. Anomalous sound-propagation phenomena

Most of our present information on winds and temperatures between about 30 km and 60 km has been derived from studies of the so-called anomalous sound-propagation from large explosions on the ground. Separate zones of audibility are frequently observed.

(i) An 'inner zone' extending outwards from the source with size and shape determined by the magnitude of the explosion and the wind and temperature distribution in the troposphere.

(ii) 'Silent zones' (where no sound is heard) at greater distances.

(iii) Further 'outer zones' of audibility at 100 km or more from the explosion point. Sound travelling to these zones often takes 1-2 min longer than if it had travelled along the earth's surface. These outer zones may exist all round the source but are frequently found only to the east of it in winter and to the west in summer. It has been shown by computing the trajectory of the sound through the atmosphere that it must usually reach an altitude of 40-60 km before returning to the earth's surface. When the explosion is very large sound can probably attain an altitude of 100 km or more before being refracted down. The absorption of sound in the atmosphere increases with its frequency and with altitude (Schrödinger 1917, Veldkamp 1951) and recordings of sound refracted down from the 50 km level are usually confined to frequencies less than 1 cycle sec^{-1} and from the 100 km level below 0·1 cycle sec^{-1} (Cox, Atanasoff, Snaveley, Beecher and Brown 1949).

For recorded frequencies in which the absorption is not large the speed of sound of a plane wave travelling horizontally is :

$$\frac{dx}{dt} = \sqrt{\frac{\gamma RT}{m}} + w \quad . \quad . \quad . \quad . \quad (7)$$

where w is the wind component along the azimuth of propagation. As the composition of the air does not vary appreciably up to the altitudes under consideration γ, R, and m may be taken as constant and

$$\frac{dx}{dt} = \text{constant}\sqrt{T} + w \quad . \quad . \quad . \quad . \quad (8)$$

If $\frac{dx}{dt}$ can be found it should therefore be possible to design experiments to obtain the temperature T and wind w. The validity of Eq. (8) has, however, been questioned by laboratory experiments (Abbey and Barlow 1948, Maulard 1949) which indicate that the speed of sound may increase at low pressures. On the other hand, there are objections to these results and no theoretical explanation of them can be given at present (Cowling 1952). Certainly the results for wind and temperature at 40-60 km obtained by assuming that Eq. (8) is valid are in agreement with those obtained by other methods. If there were a change in the nature of propagation at low pressure, e.g. due to the disturbance no longer being small compared to the ambient pressure, it would only affect the equation for the speed of sound in the air and the component w representing the air movement derived from Eq. (8) would still be correct.

The experimental procedure is briefly as follows. The time taken by the sound to travel a known distance from the explosion point to a lattice of microphones is measured. The apparent speed and direction of the sound travel over the lattice are found from the times of arrival at the individual microphones. From these data and wind and temperatures from a simultaneous radiosonde, the trajectory of the sound and its time of travel up to the top of the balloon ascent can be calculated. The distance and time at higher levels are then found. If some law of variation of wind and temperature with height is assumed, solutions for the speed of sound at different altitudes can be obtained. If simultaneous measurements of this type are available on three or more azimuths, values of wind and temperature may then be found from Eq. (8).

Whipple (1935) suggested that high temperatures in the ozonosphere and a monsoon system of winds (easterly in summer and westerly in winter) were necessary to explain

the observed audibility features. Wölcken (1934) obtained evidence that these high temperatures also exist during the polar night. Generally, however, the early experimenters were not able to obtain sufficient measurements to separate the wind and temperature effects (Gutenberg 1951). More recently Murgatroyd (1956a) in England and Crary *et al.* (1950, 1952, 1953) in the U.S.A., using more elaborate facilities, have devised methods of doing this and given results for winds and temperatures up to 50 or 60 km. Both methods assume simple laws for the variation of wind and temperature or speed of sound with height and hence the results are incorrect in detail. As shown in Fig. 3 and 4, however, they all fall into a consistent framework and probably give a good general picture of wind and temperature conditions at these levels. Murgatroyd concluded that over England between 20 and 30 km there was a small lapse of temperature in winter and isothermal conditions in summer. At higher levels there was an inversion in both seasons and the temperature between 40 and 50 km approached that at the surface. There was a seasonal variation of temperature of about 40°C at 50 km. The wind at this level was westerly 50-60 m sec^{-1} in winter and easterly 20-30 m sec^{-1} in summer. Zonal wind components were usually considerably greater than meridional components. Crary *et al.* have reported experiments for different seasons of the year in various parts of the United States, Alaska, the Panama Canal Zone and Bermuda. Other studies in the U.S.A. using Crary's method have been made by Richardson and Kennedy (1952) and Kennedy, Brogan and Sible (1955). The temperature structure obtained was broadly similar in all cases with about 200°-220°K around 20 km and an increase with altitude to values roughly equal to surface temperatures between 40 and 50 km. The seasonal variation at high levels increased with latitude. In Alaska the temperature at 50 km was 290°K in summer but only about 240°K in winter. The height of the maximum of temperature also increased with latitude, 260°K at 60-65 km being reported for Alaska in winter. The corresponding winds varied with season and latitude, generally being east in summer and west in winter. In low latitudes speeds were greatest in summer, while at high latitudes they were greatest in winter, 60-70 m sec^{-1} being reached at 50 km or so in both seasons. Strong southerly winds were also noted in the Alaska results. According to Kennedy *et al.* (1955), short-term (less than an hour) wind fluctuations similar to those in the troposphere occur in the 30-60 km layer, their magnitude being of the same order as the wind vector itself. This result must, however, be accepted with the reservation that tropospheric variations would affect the results of the sound recordings. Some day-to-day variation is likely as evidenced by different results of explosions on different dates but at the same time. Diurnal changes were also noted, maximum temperatures and minimum winds occurring in the mid-afternoon. This result is in agreement with the calculations of Johnson (1953) who estimated several degrees' variation due to absorption of solar radiation by ozone.

Another interesting series of experiments has been conducted by Johnson and Hale (1953) in the south-western United States. Large explosive charges of 1,200 lb T.N.T. were used with microphones on east-west lines 175 km long. During winter, propagation to the west was only possible by sound waves which attained 70-100 km (which had longer travel times and greater characteristic velocities than those from the lower inversion) whereas in the summer they were refracted down from 40 km or so. Similarly propagation to the east occurred mainly by refraction down from the higher inversion in summer and from the lower inversion in winter. Returns from the upper inversion have also been reported by Cox *et al.* (1949) and Crary and Bushnell (1955). The latter, working in October, when winds at 50 km are probably at a minimum, have made the first attempts to use the returns from about the 100 km level to estimate wind and temperature there. Their records, however, were not extensive enough to give detailed data.

3.5. *High-altitude rockets*

The development of rockets capable of carrying instruments to several hundred kilometres above the earth's surface has made it possible to obtain direct measurements of pressure, density, and wind as well as to acquire new data on solar radiation, atmospheric composition, the ionosphere, the earth's magnetic field and cosmic rays. Reviews of the methods of measurement together with a selection of results have been given by Newell (1953) and Boyd and Seaton (Eds.) (1954). The instrumentation used for this work is complex and involves telemetering the results from rocket to surface or the recovery of recording equipment after flight. The measuring elements all have special features designed to overcome the conditions imposed by rocket flight, e.g., they must be robust, have adequate exposure and very rapid time of response as well as being sufficiently accurate and sensitive. The trajectory of the rocket is obtained by optical or radar tracking and hence its velocity and height at any instant can be found. Supplementary data on its behaviour in flight such as angle of attack, azimuth and attitude of the rocket axis, yaw and roll may be obtained from indicating vanes or gyroscopes. In addition to their application to performance studies many of these data are necessary in the derivation of pressure, density and temperature. Recently Reisig (1956) has described how the wind can also be determined from measurements of the velocity vector, attitude and angle of attack, and has presented preliminary results.

The direct measurement of temperature by some form of thermometer element is extremely difficult owing to the large conduction and kinetic heating effects and the small lags required, and it is usually derived from measurements of pressure on the rocket surface. The instruments used for the latter purpose are the bellows gauge from about 1000-20 mb, the Pirani gauge from about 2 to $2{\cdot}10^{-3}$ mb, the Phillips gauge from $2{\cdot}10^{-3}$ to 10^{-6} mb and the Havens cycle gauge from 1000 to 10^{-3} mb. If p_x is the measured surface pressure at a distance x from the nose of the rocket and p is the ambient pressure then (Newell 1953) :

$$\frac{p_x}{p} = f(x, T, v, \alpha, \theta, m, \gamma) \quad . \quad . \quad . \quad . \quad (9)$$

where

T = ambient temperature

v = rocket speed

α = angle of attack

θ = angle of roll

m = mean molecular weight

γ = ratio of specific heats

The last two variables involve a knowledge of atmospheric composition which can be taken as invariant up to 70 or 80 km and then is affected by the gradual change from molecular to atomic oxygen at higher altitudes (Jones 1954, Chackett *et al.* 1950). The problem is to design an experiment which will allow p, T or atmosphere density ρ to be found from one or more measurements of p_x, v, α and θ, using either aerodynamic theory or comparison with wind-tunnel tests. Among the principal methods used up to date may be mentioned the following:

(*a*) It has been found by wind-tunnel tests that, for small angles of attack, the pressure on the rocket surface towards its rear is very nearly equal to the ambient (static) pressure so that in this case p can be obtained directly (Havens, Koll and Taylor 1952). To correct for yaw, similar gauges are installed on opposite sides of the rocket and their readings averaged.

(*b*) Pressures measured on the surface of a conical nose can be related by the Taylor-Maccoll theory to values of $\frac{p_x - p}{\frac{1}{2}\rho v^2}$, so that if p_x is measured at two or more points both p and ρ can be calculated (Sincinski, Spencer and Dow 1954).

(*c*) Below about 100 km the ram pressure p_r measured by a forward-facing pitot tube can be interpreted by the Rayleigh supersonic pitot-tube formula in the form

$$\rho = \text{constant}\,(p_r/v^2) \quad . \quad . \quad . \quad . \quad . \quad (10)$$

The constant depends on atmospheric composition and the formula is accurate only at Mach Numbers greater than about three. At higher altitudes where the molecular mean free paths are too long for (10) to be valid, a kinetic theory due to Havens *et al.* (1952) has been used.

(*d*) Deductions of atmospheric pressure or density at higher levels, where outgassing from the rocket creates a back pressure which invalidates other pressure measurements, may be possible when a rocket is deliberately rolled or yawed. The amplitude of the pressure modulation measured on the side of a rolling rocket is directly proportional to the product of the ambient density and the component of the relative air velocity perpendicular to the gauge opening (Havens *et al.* 1952).

When the pressure at a given height is known the density can be found from the slope of the pressure-height curve using the hydrostatic equation. Similarly temperature can be derived from the slope of the log pressure-height curve although this requires a knowledge of atmospheric composition. The process of differentiation required may, in both cases, lead to considerable errors.

Other methods of deriving the temperature involve a measurement of the local Mach Number M and hence the speed of sound, where

$$M = \frac{v}{a} \quad . \quad . \quad . \quad . \quad . \quad (11)$$

This again requires a knowledge of atmospheric composition. The shock-wave angle

$$\mu = \sin^{-1}\frac{1}{M} \quad . \quad . \quad . \quad . \quad . \quad (12)$$

can be measured in flight by direct photography of the wave by Schlieren technique. Another system described by Bartman (1954) requires radar tracking of a sphere ejected from the rocket in flight, in order to measure the drag D on it as it falls and hence the ambient density.

$$D = \tfrac{1}{2}\rho v_1^2 c_D A \quad . \quad . \quad . \quad . \quad (13)$$

where v_1 is the fall speed of the sphere, A its projected cross-sectional area, and c_D its drag coefficient previously determined by calculation or wind-tunnel measurements.

Finally measurements of the speed of sound and hence temperature and also wind have been made in the rocket grenade experiments reported by Stroud, Terhune, Venner, Walsh and Weiland (1955), Ference, Stroud, Walsh and Weisner (1956), and Weisner (1956). Theoretical aspects of this work have also been discussed by Groves (1956). Grenades containing a small explosive charge were ejected from rockets at altitude intervals of 5 to 10 km and exploded. The time and position of these explosions were found by means of a ballistic camera and the time of travel of the sound down to a lattice of microphones roughly vertically below was recorded. Use of the ballistic camera and the requirements of good conditions for sound reception have so far restricted the experiments to occasions of clear nights. As it travels from explosion point to microphone

the sound is refracted in its path by changes of wind and temperature. Enough data are obtained to separate the two effects and deduce the mean wind and temperature in the layers between successive bursts. The only uncertainties involved arise from relatively minor mathematical assumptions and the conversion of sound speed to temperature discussed above in § 3·4.

Although this outline of the various experiments includes those which have given most of the results published up to date, it is not exhaustive. For instance X-ray absorption measurements are a possible method of determining pressure at high altitudes (Newell 1953). A summary of the pressure, density, and temperature measurements up to 1952 has been given by the Rocket Panel (1952) and a selection of these are included in Fig. 3 The average temperatures broadly confirm the values obtained by other methods and show a maximum of about 270°K at 45-50 km, and a minimum of 200°K around 80 km with an inversion above. The great majority of the firings have so far been at White Sands, New Mexico, latitude 32°N. Although the measured values varied from firing to firing, with significant changes apparently occurring within a few hours, the agreement between the mean temperature results of the principal methods has been within 10 or 20°K at all altitudes. It has not been possible to detect any regular seasonal or diurnal variation from the values obtained up to date and it seems that at this latitude they do not exceed 10-15°K at any level. The wind measurements from the rocket trajectory calculations and the grenade experiments show that the wind in the ozonosphere is mainly easterly in summer and westerly in winter up to an altitude of 60-70 km. Speeds are high in both seasons and reach a maximum of 80-100 m sec^{-1} between 50 and 60 km decreasing to considerably smaller values between 60 and 70 km. Representative values are included in Fig. 4. The calculations of Reisig (1956) from trajectory data in three firings in Florida indicate considerable fine structure and variability of wind speed and direction with height in the 30-60 km altitude band. So far there is little information on conditions at other latitudes. Apart from a few flights near the equator which have given temperature results very similar to those obtained at 32°N the only results so far published are those for two summer firings in the Arctic using ' Deacon ' rockets released at altitude from ' Sky-hook ' balloons (LaGow and Ainsworth 1956). In both of these very high temperatures (320-350°K) at about 40 km were reported.

3.6. Searchlight measurements

It is possible on a sufficiently clear night to obtain a measurement of density ρ by measuring the intensity E of light scattered at an angle β down from the beam of a high-intensity searchlight. Assuming Rayleigh scattering

$$E = \text{constant}\ (1 + \cos^2 \beta)\, \rho \qquad . \quad . \quad . \quad (14)$$

The constant in Eq. (14) can be found by a comparison with radio-sonde observations made at the same time. The method can be used for altitudes from about 10 km to 70 km at present, the lower limit being determined by the extraneous scattering due to dust in the troposphere and the upper limit by the decreasing contrast with the light of the night sky. The beam may be modulated to improve the latter, and detected with a photo-multiplier situated about 20 km from the searchlight (Elterman 1954). This is a very useful method of obtaining upper-air densities at night time, once the technical difficulties have been overcome, because of the comparative simplicity of the apparatus used and the ease of observation. Recently an attractive system using only one station and employing pulse techniques has been developed (Friedland, Katzenstein and Zatwick 1956) and results similar to those obtained with the two-station system have been reported.

Elterman's density measurements obtained over New Mexico agree well with the rocket results. The temperature at the various altitudes can be derived from the densities by using the hydrostatic equation and equation of state, provided a value obtained from radiosonde results is used as a starting point at the lowest altitude. Unfortunately, however, errors are additive so that above 50-60 km the calculated temperatures are not very reliable. The results show a maximum at 53 km where the value is between 280° and 310°K. Little seasonal trend has been found below 50 km but at greater altitudes temperatures were 20°-30°K higher in early autumn than in mid-summer. Indications were also obtained of up to 10°K temperature decrease throughout the night.

3.7. Visual and photographic observations of meteors

Meteoric particles move in orbits round the sun and are attracted into the earth's atmosphere when it passes sufficiently close to these orbits. It has been estimated (Mitra 1948, p. 73) that about 10^9 meteors enter the earth's atmosphere daily. They commonly have speeds relative to the earth of some tens of kilometres per second and, as they collide with the air molecules, the heat developed vaporises them and often makes them visible to the naked eye. Their positions and times of appearance, disappearance and variation of luminosity have frequently been observed visually or by photographic methods. It is possible to use these data to estimate air temperacure in the 50-100 km altitude band. In addition observations of the drift of meteor trails have produced information on winds at these levels.

The first comprehensive theory of the vaporisation of meteors was developed by Lindemann and Dobson (1923). They assumed that a hot-gas cap forms in front of the meteoric particle owing to the adiabatic compression of the air and that heat from the gas cap is then transmitted to the meteor which finally becomes incandescent. Formulae were developed for the density required for the formation of the gas cap, the rate of loss of energy by the meteor and the energy available for heating it. The meteor mass was estimated by an integration of the light-intensity curve. Using these concepts it was possible to calculate separately the air density at the height of appearance and disappearance. The results showed that the air densities above 60 km were in the order of 10^3 times higher than would have been found if the atmosphere from 20 to 60 km remained throughout at the temperature of the lower stratosphere (220°-230°K). These high densities could only be explained by a considerably higher temperature – *circa* 300°K – around 50 km and this was the first evidence that this temperature maximum exists. Other supporting evidence was put forward based on the fact that around 55 km there is a drop in the frequency of disappearance of meteors which was attributed to the presence of an isothermal region of temperature maximum there. Although some aspects of the theory have been criticised (e.g., Sparrow 1926) there is little doubt about the main conclusions and it has provided a basis for subsequent work. Öpik (1936, 1937) has examined large amounts of data obtained visually in Arizona and put forward a theory of the vaporisation process. Photographic techniques have mainly been developed at Harvard Observatory (Whipple F. L. 1940, 1943). Simultaneous photographs against a reference background of stars were obtained using cameras at the end of a 40 km base line. The data comprised the trajectory, velocities, deceleration and the light curve. A composite theory was developed utilising the same physical data and concepts as Öpik and a similar mathematical treatment to that of Hoppe (1937). The luminosity measurements combined with the velocities allowed the rate of depletion of the meteor's mass along its track to be found, and the deceleration was then used to find the drag and hence the atmospheric density at each point. The method of evaluating the Harvard data has since been modified slightly by Jacchia

(1948, 1949) who gave mean temperatures of 281°K at 70 km, 224°K at 80 km, 235°K at 90 km, 255°K at 100 km and 256°K at 110 km. These temperatures were found by integrating the measured density-height curve to obtain the pressure at each height, and then using the equation of state, assuming a suitable value for the mean molecular weight of the atmosphere at the level concerned.

Although the progress made so far with visual studies of meteors has been impressive it seems doubtful whether these various techniques can be made to give more than very general information on upper-air densities and temperatures. There are uncertainties in some of the data used (e.g. shape factor, drag coefficient, heat-transfer coefficients and luminous efficiency factor) and limitations of this type have made it advisable to anchor the density curve produced from the meteor equations to the density obtained by other means in the 50-60 km zone.

The investigation of winds by visual and photographic observations of visible trails has also considerable limitations. These arise in part from the presence of large tidal components in the winds above about 70 km. To separate these from any seasonal winds, in addition to discovering any variation with latitude and altitude, requires many more measurements than have been made since wind drifts from careful observations of the movement of visual meteors have been reported. Olivier (1942, 1947) has listed about 1,600 reports mainly from America, Western Europe and Russia. The author (unpublished), doubtless in common with many others (e.g., Chiu 1951), has attempted to extract broad conclusions from them. Chiu reported he could find no well-marked trends with height. The tentative conclusions of the present author, based on frequency tables of wind direction at different times of the day, were :

(a) Winds at 80-100 km are predominantly westerly in summer, with easterly winds the least frequent.

(b) They are also predominantly westerly in winter, with northerly winds the least frequent.

Other results from visual studies of drifts are :

(i) Wind speeds between 50-100 m sec^{-1} are frequent between 70 and 100 km, and 30-50 m sec^{-1} between 30 and 70 km. Occasionally considerably higher speeds have been reported.

(ii) Some evidence has been advanced for significant changes at the 80 km level, Kahlke (1921) suggesting that wind directions are generally easterly below it and westerly above. Trowbridge (1907), Hoffmeister (1946) and Fedynsky (1955) obtained results substantially in agreement.

(iii) There is evidence of large vertical shears with different wind directions at different levels and considerable turbulence at all levels from 40-110 km.

Although these results do not lead to any definite conclusions about the wind systems of the atmosphere, they are a useful guide to speeds and the directions at least do not contradict the results of the other studies. Some very useful information on turbulence has been given by the photographic studies of the Harvard group (Liller and Whipple 1954) and vertical wind shears of up to 50 m sec^{-1} km^{-1} have been reported.

3.8. Radio observations of meteors

Meteors can be tracked by radar and Doppler methods using pulse or continuous wave techniques and hence measurements of their heights, velocities and decelerations can be made. Radio reflections can be obtained in the 4-10 m band from the ionised column they produce (Hey 1949, Lovell 1946). Temperature can be deduced from observation of the trajectories and wind from the measured movement of the ionised columns.

In one method of temperature measurement developed by workers at Jodrell Bank, England (Evans 1955) height distributions of large numbers of sporadic meteors in given velocity groups are first obtained. For a given velocity group the r.m.s. deviation in a height distribution depends only on the mass distribution of the meteors and the scale height H (Lovell 1956). The differential mass distribution has been given by Kaiser (1954) in the form

$$dN \propto M^{-2}\, dM \quad . \quad . \quad . \quad . \quad (15)$$

where dN is the number of meteors in the mass group M to $M + dM$. Hence for each velocity group the scale height H can be determined and its variation with altitude from about 85 km to 110 km has been found (Evans 1954). Temperature can be directly found from the scale height if the composition is assumed, and pressure is obtained by integrating the scale-height profile. The variation of pressure with height over the same altitude range has also been determined by measuring the mean heights of the reflection points for meteors in the different velocity groups and applying a formula given by Herlofson (1948) for the atmospheric pressure at the height of maximum ionisation. The latter is determined by a method due to Kaiser (1954) in which the maximum ionisation of the smallest detectable meteor is related to the observed mean height of meteors in that velocity group.

The temperature and density results obtained at Jodrell Bank (Evans 1954, 1955) show good agreement with those obtained in New Mexico by rocket and photographic meteor techniques. At 85 km the scale height was found to be approximately 6·0 km and at 105 km 7·5 km. The corresponding values of $\log_{10}$ density (g cm^{-3}) were $-$ 8·0 and $-$ 9·2. Assuming a mean molecular weight of 29, the respective temperatures would be about 200° and 250°K. Although there are uncertainties in the procedures described above, these results are very encouraging and very probably further advances in the techniques can be expected. According to Lovell (1956) the prospects are extremely good – ' the technique (of obtaining scale heights and variation of pressure with height by radio-meteor observations) therefore opens up the prospect of establishing the extent of any diurnal or seasonal variation and of investigating any latitude dependence.'

The development of techniques for wind determination using radio reflections from the ionised column has led to detailed information on mean winds, seasonal and diurnal effects in the 80-100 km altitude band. It is unlikely that the observed drifts are due to any other effect than true wind motion since the reflecting columns have electron densities considerably greater than those of the surrounding atmosphere. The electrons, which cause the echoes, are held in the narrow column by strong electrostatic forces and the column drifts as a whole (Greenhow 1954). Manning, Villard and Peterson (1950) in the U.S.A. have used a continuous wave Doppler method together with a special direction finder. Greenhow (1952) at Jodrell Bank used a coherent-pulse Doppler technique, and Robertson, Liddy and Elford (1953) in Australia used a combination of pulse and continuous wave methods. Since reflections can be recorded both by day and night independently of cloud it is possible to accumulate data very rapidly. Each system involves the measurement of the radial motion of the ionised column relative to a station. Reflections from a minimum of two trails at different measured ranges, azimuths and elevations from a station give the minimum two components which are sufficient to determine the magnitude and direction of the movement. Greenhow used a beamed aerial directed alternately for about 10 min in two directions at right angles to obtain average perpendicular components of velocity. Measurements may be made for about $\frac{1}{2}$ hr to evaluate the mean motion, usable individual echoes often being obtained at the rate of more than one per minute and recorded by cine-photography of the cathode-ray tube displays. Using the pulse technique it is possible to obtain over 3,000 usable echoes each day. The area

investigated may be about 200 km long, 100 km broad and the depth 10 km or so, the mean height of investigation usually being about 95 km. Techniques have also been developed to find the variation of wind with height and the degree of turbulence (Lovell 1956). The former is only possible when the echo rate is high as most meteor observations are obtained in a narrow height band. In addition, calculations have been made of diffusion coefficients from a study of the rate of decay of the ionized columns. Results in England (Greenhow and Neufeld 1955) and also in Australia give the value of 10^4 cm^2 sec^{-1} at 80 km and 10^5 cm^2 sec^{-1} at 100 km. These agree with theoretical values for molecular diffusion obtained from the Rocket Panel's estimation of density at these levels.

The results of the wind determinations may be summarized :

(i) A large semi-diurnal component is always found. Over England its amplitude increases with height from about 10 m sec^{-1} at 85 km to 15 m sec^{-1} at 100 km in summer. There is a corresponding increase in winter from 10 m sec^{-1} to 30 m sec^{-1}. The phase is very variable and complex changes occur in the autumn. There is an advance of phase with height of about 3° km^{-1} in summer and 7° km^{-1} in winter. On the average the component is southerly near 0600 and 1800 hr at 100 km but there are considerable seasonal variations. Details have been given by Greenhow and Neufeld (1956). These observations are in satisfactory agreement with the resonance theory of atmospheric oscillations. Moreover the results of Elford and Robertson (1953) in Australia are also in substantial agreement, the components there being 180° out of phase with those in the Northern Hemisphere, as predicted.

(ii) There is also a diurnal periodic wind of amplitude about 5 m sec^{-1} which appears mainly to be southerly at about 1200 hr over England and to have little variation with height or season. In Australia, however, the diurnal component is considerably larger and is southerly at 1800 hr.

(iii) Other periodic components are insignificant.

(iv) The prevailing wind in England at 85 km has westerly components of about 20 m sec^{-1} in winter and 10 m sec^{-1} in summer; at 100 km they are westerly 15-20 m sec^{-1} in both seasons. In spring easterly components of 10 m sec^{-1} are found at both 85 and 100 km. In autumn there is a westerly component at 85 km and an easterly component at 100 km both being about 10 m sec^{-1}. The overall trend is for westerly winds to increase with height in summer and early winter but decrease with height in the autumn and late winter. An extrapolation to 110 km would suggest westerly winds at this level in summer and possibly easterly in winter. Similarly below 85 km winds tend to be more easterly in summer and westerly in winter. There are northerly components of 10-15 m sec^{-1} at both 85 and 100 km in summer and smaller southerly components are more common in the other seasons. The prevailing wind shows irregular day-to-day variations and changes in direction; as much as 90° change has been observed over a period of 12 hr. The results of Manning, Villard and Peterson (1950) for a number of occasions during the summer in the U.S.A. are in reasonable agreement with the foregoing. In Australia in summer between 80 and 100 km westerly winds increasing with height are found. In winter, however, although winds were again westerly, they decreased with height, possibly reversing the direction at about 100 km. Winds were westerly in autumn but easterly in spring (Weiss 1955). These results are in good agreement with those for England.

(v) Turbulent winds with r.m.s. velocities of 30 m sec^{-1} are always observed. Wind shears of 50 m sec^{-1} over height differences of a few kilometres have been found. Irregularities in the wind structure down to a scale of 500-100 m have been observed.

(vi) The wind is substantially horizontal but individual observations may give up to 10 m sec^{-1} vertical components.

(vii) It also appears probable from Greenhow and Neufeld's (1956) results that the changeover from west winds to east winds and vice versa occurs a month or so earlier at the higher levels (90-100 km) than it does at lower levels (80-90 km) but further evidence is required on this point.

3.9. Noctilucent clouds

Luminous night clouds at a height of about 80 km have been observed from time to time from high-latitude stations during the summer. The majority of the reports come from Norway (Störmer 1935), Canada (Vestine 1934), Scotland (Paton 1949, 1954) and Russia (Khvostikov 1952). Their altitudes and drifts have been measured by triangulation methods using visual or photographic observations. They are visible only by scattered light from the sun when it lies 10°-18° below the horizon and they have rarely been seen higher than 10° above the horizon. According to Vestine (1934) conditions of illumination for them to be observed by scattered sunlight can occur at other latitudes often enough for them to be seen if they existed. The only observation, however, of their possible existence at lower latitudes is a report by Miley, Cullington and Bedinger (1953) of cloud photographs recorded by a rocket-borne camera at about 80 km at latitude 33°N.

It is not known for certain whether these clouds are composed of ice or dust of either volcanic or meteoric origin. The first two require the existence of convective exchange from the earth's surface to their level. It is interesting that the first suggestions of a low temperature at 80 km, now confirmed by rocket experiments and other work, came as a result of calculations by Humphreys (1933) of the temperature necessary for ice saturation there. There is considerable doubt about the validity of Humphreys's assumptions, however, and Ludlam (1955) has advanced reasons based on optical considerations why these clouds are more likely to be dust. A temperature minimum near 80 km would probably be associated with an accumulation of dust there but nothing can be deduced about its absolute value.

The measured winds vary considerably but in general they have the following common features.

(a) The general drift is predominantly to the west indicating mainly easterly winds, although large north and south components are also found.

(b) Measured speeds vary between 20 and over 100 m sec^{-1}, the mean being 40-50 m sec^{-1}.

(c) There appears to be a clockwise rotation of the wind vector throughout the night. Jesse (1896) stated that before midnight most movements were from an azimuth of 48°E and after midnight 63°E, with a possible secondary maximum from 73°W. Vestine (1934) gave a diagram which suggests that the wind is mainly northerly about 2100 hr and easterly at 0300 hr local time. A mean easterly wind with a superimposed semi-diurnal component could account for this result although other explanations are possible. In the construction of Fig. 4 an easterly component of 40 m sec^{-1} has been used as the mean wind at 80 km in summer at 70°N as a representative value based on this information.

(d) There are some dubious features in several of the quoted observations. In particular persistent apparent movements of the clouds away from the sun and increased velocity towards the end of the observing period suggest some confusion of wind drifts with optical effects in the observations.

(e) Finally, turbulent effects and wavelike motions in the shapes of the cloud have been reported frequently. This observation is in agreement with visual observations of long-enduring meteor trails at similar altitudes.

3.10. Radio propagation studies

Studies of the propagation of electromagnetic waves through the ionised layers of the earth's atmosphere have produced information on both temperature (mean kinetic energy of the gas particles) and wind at levels from about 60-500 km. The presence of free electrons alters the index of refraction and hence the phase velocity of the radio waves, and collisions of ionised and neutral particles produce dispersion. The ionization is mainly caused by absorption of the shortest wave bands of solar radiation and some recombination occurs during the night. Although there is a background level of ionisation at all levels in the upper atmosphere, layers of maximum concentrations have been observed as follows :

TABLE 3. PRINCIPAL IONIZED LAYERS.

Layer	Height (km)	Electron densities cm^{-3}
D	60-100	$1{\cdot}5 \times 10^4$
E	90-120	$1{\cdot}5 \times 10^5$
F_1	160-220	$2{\cdot}5 \times 10^5$
F_2	250-500	$1{\cdot}5 \times 10^6$

A high electron density of about $3{\cdot}0 \times 10^6$ electrons cm^{-3} known as ' sporadic E ' often exists locally in patches at about 120 km. There may also be a ' G ' layer at about 600 km.

The methods of applying propagation theory to utilize radio measurements for determination of the temperature in the ionosphere, and the results up to date, have been reviewed by Gerson (1951) and by Kaplan, Schilling and Kallmann (1955). The principal methods are :

(i) Measurements are made of the apparent height of reflection as a function of frequency. The variation of free-electron concentration with height can then be determined. Theories for the formation of ionised layers show that the semi-thickness of the layer (the vertical distance between zero and maximum electron concentration) is approximately equal to twice the scale height H. The scale height can also be estimated from the variation of height of maximum electron density with solar zenith angle.

(ii) A measurement of the variation with frequency of the reflection coefficient leads to values of the electron-collision frequency. Alternatively it can be estimated from studies of the cross modulation of two radio channels. The collision frequency thus determined depends on the effective collision cross-sections of the electrons, positive ions and neutral particles and the relative concentration of the various constituents at the level concerned as well as the temperatures there.

(iii) Studies are made of recombination coefficients. These vary both with temperature and composition.

The results of all these investigations require a knowledge of atmospheric composition before the temperature can be determined and they are also subject to many assumptions arising from the theory of formation of the ionised layers and the values of the ionospheric parameters used. Budden, Ratcliffe and Wilkes (1939) deduced a scale height of about 6 km – equivalent to about 200°K – for the 70-80 km region. Most of the other reported results refer to the E and F regions and show scale heights increasing considerably with altitude. The uncertainty of interpreting these latter results in terms of temperature is considerable. The most recent estimate of mean values above 100 km making reasonable assumptions about atmospheric composition has been given by Kallmann, White and Newell (1956) who concluded that the temperature increases from 237°K at 100 km to

900°K at 300 km. The corresponding pressures at these levels are $5{\cdot}7 \times 10^{-1}$ and $1{\cdot}7 \times 10^{-5}$ dynes cm^{-2} and the densities $8{\cdot}3 \times 10^{-10}$ and $4{\cdot}4 \times 10^{-15}$ g cm^{-3} respectively. The measurements are at present regarded as too uncertain for diurnal, seasonal or latitudinal variations to be deduced from them and the above results should be regarded as general means.

The determination of wind from radio-propagation measurements is only possible when there are horizontal irregularities in the ionisation whose apparent movement can be tracked. The question arises as to whether the detected motion is a wind or only a movement of charged particles relative to the atmosphere as a whole. Fejer (1953), who has made calculations of the tidal motions in the ionosphere necessary to produce the observed quiet-day magnetic variations, concluded from considerations of their phase that fading measurements of movement in the 'E' layer *c.* 100 km must refer to air motions and not electron drifts. The problem has also been considered by Clemmow, Johnson and Weekes (1953) and Stolov (1955) and it appears probable from their results that the movement of charged particles detected by radio methods can be interpreted as a wind up to 100-150 km. There are three main techniques based on the following :

(a) fading records (random amplitude variations of received signal strength), (b) the movement of patches of 'sporadic E,' and (c) the amplitude variations of signals received from radio stars.

The principal method for the E layer has been to compare the fading records obtained from three receivers with their aerials situated less than one wavelength apart at the corners of a triangle. These records are generally similar (unless there is a large amount of turbulence) and they arise from the drift over the ground of an amplitude pattern. The direction of movement is the same and the speed twice that of the corresponding scattering centre in the ionosphere. The techniques, different methods of analysis and results up to date have been reviewed by Briggs and Spencer (1954). Mayaguez has also described the various methods and given a summary of other references. Using frequencies of about 2 Mc sec^{-1} results have been obtained for the E layer by day and the F layer by night. They are in sufficient detail, in the case of the E layer, to allow diurnal and seasonal components to be resolved. Lower frequencies *c.* 100 Kc sec^{-1} will give E layer reflections by day and night. There is an uncertainty in the precise height at which the reflection occurs but it is usually considered to be around 100 km. In other rather similar methods, used mainly for F layer work, observations of the movements of larger-scale irregularities have been made by using transmitters or receivers at widely spaced points (e.g., Wormell 1954, Obayashi 1956, Briggs and Spencer 1954).

Sporadic E layer ionisation becomes evident when low-power radio communication is possible at frequencies considerably greater than the normal maximum usable frequency between two given points on the earth's surface. If the skip distance changes with time so that other pairs of points in turn establish communication it can be concluded that there is a dense patch of ionisation drifting with a wind. Ferrell (1948) and Gerson (1955), with the co-operation of large numbers of radio amateur operators in North America, have been able to obtain measurements of the motion, probably at about 120 km, by this method. Other wind measurements from sporadic E receptions at 5-8 Mc sec^{-1} using a central receiver and spaced transmitters in the day time have been reported from Australia (Harvey 1955), and they have also been possible at night in this country using the equipment normally used for daytime E layer recordings by the fading technique.

The variations in intensity of signals received from radio-stars are thought to be due to passage through irregularities which move in the ionosphere, probably in the F layer (Maxwell and Little 1952). It is, however, very doubtful whether the drifts calculated from this work can be interpreted as winds.

The results of the various workers who used fading records suggest a mean velocity of about 50 m sec^{-1} in the E layer with a tendency for somewhat greater velocities in winter than in summer. Briggs and Spencer (1954) by taking the wind direction at 1200 hr each day in their measurements at Cambridge, England, also found that the zonal winds there were westerly in summer and easterly in winter, i.e., opposite in direction to those in the ozonosphere. Previous British measurements (Phillips 1952) and results from the U.S.A. (Salzberg and Greenstone 1951) and New Zealand were in substantial agreement with this conclusion. Harang and Pedersen (1956) obtained similar results in Norway. Greenhow and Neufeld (1955), however, criticised Briggs and Spencer's analysis because results taken at 1200 hr did not always eliminate the tidal effect. Re-analysing the data by plotting the zonal component each day at the mean time of zero tidal wind components, 1400-1500 hr, they came to the conclusion that Briggs and Spencer's data gave a seasonal variation of mean zonal wind component in good agreement with that deduced from their own radar-meteor observations. The prevailing zonal wind component in the E layer over England is then westerly in summer and winter with speeds of 40-50 m sec^{-1} and easterly 10 m sec^{-1} in spring and 20 m sec^{-1} in autumn. Compared with the meteor results this indicated little change in the easterly components at the equinoxes but an increase of the westerly components at both solstices – perhaps corresponding to a greater altitude of the fading records compared with the meteor records. These results are not entirely in agreement with those of the radar-meteor observations at Adelaide, (Weiss 1955) or at Jodrell Bank (Greenhow and Neufeld 1956) which suggest that the west wind increases with height in the summer but decreases in the late winter, tending to become easterly about 100 km. The meridional components of wind showed no systematic variations. In England semi-diurnal variation was also found equivalent to a wind vector of about 30 m sec^{-1} rotating clockwise and southerly in direction at 0300 and 1500 hr. Its magnitude was smaller in summer than in winter and there was some evidence of a seasonal variation in the phase. In the southern hemisphere results were somewhat similar, the rotation being anticlockwise but the phase displaced about two hours. Further evidence that the phase of the semi-diurnal component varies with altitude and season has been produced by I. L. Jones, who used closely spaced aerials and frequencies of 2 and 2·5 Mc sec^{-1} to obtain reflections at points about 5 km vertically displaced in the E layer (Ratcliffe 1956). A phase advance of 5°-10° km^{-1} would explain why the fading-record results of Briggs and Spencer (which give a southerly vector at 0300 hr) compared with those of Greenhow and Neufeld (which give a southerly vector at 0600 hr) are 90° advanced in phase, if the fading records referred to an altitude of about 10 km higher than the meteor results. However, in view of the seasonal changes of phase reported by both the workers concerned with radar meteor and fading-record studies it appears likely that the vertical temperature distribution will also be important in determining the phase. Little information on the 24 hr diurnal component has been obtained from the technique of using fading records.

The two years' series of measurement of daytime drifts of sporadic E layer ionisation by Harvey (1955) in Australia gave directions of movement generally opposite to those obtained from the meteor observations of Elford and Robertson (1953). The mean wind velocity was about 60 m sec^{-1} and the directions roughly south-easterly in both summer and winter. The observed change in wind direction could simply be due to the two sets of observations being at different altitudes, the sporadic E irregularities perhaps producing reflections from altitudes up to 20 km higher than the meteors. Gerson's results for North America also show mainly easterly winds particularly in the early summer.

Summarizing, it appears that the main conclusions from the radio propagation observations in the region of the E layer are :

(i) In both summer and winter mean zonal winds are westerly in the 80-100 km altitude band. In spring and autumn they are easterly and of smaller magnitude.

(ii) Speeds are about 50 m sec^{-1}.

(iii) Meridional components are often as strong as zonal components but are not so regular, and turbulence appears to be a feature of all reported observations.

(iv) Over England, speeds increase with height in summer and winter. At lower latitudes they decrease with height and winds probably become easterly above 100 km in winter.

(v) Semi-diurnal components are about the same magnitude as mean components. They tend to increase in magnitude and advance in phase with height.

(vi) There is a tendency for easterly wind components to prevail towards the top of the E layer.

Considerably more observations throughout the day over at least a year and referring to known altitudes, are required in different latitudes before a satisfactory picture of the general circulation in the 80-120 km altitude band can be obtained. However, the experimental methods to do this exist and it seems likely that this region can be explored more easily than the 40-80 km altitude band.

The observations of the motion of ionised patches in the F layer have revealed the following features :

(i) Speeds are higher than in the E layer. 100-150 m sec^{-1} is a common value in the F layer and considerably greater values have been reported under magnetically disturbed conditions.

(ii) The zonal component is westerly by day and easterly by night.

(iii) The meridional component is southerly in summer and northerly in winter in the northern hemisphere with opposite directions in the southern hemisphere.

There are not enough data to interpret these observations in terms of mean wind systems and oscillatory components nor, indeed, can the motions be interpreted in terms of winds until the details of the magnetic-hydrodynamic control are better understood.

3.11. The airglow and the aurora

The light received at the earth's surface on a moonless night arises from several sources – stellar, zodiacal light, luminescence of the night sky (airglow) and, mainly in high latitudes, the aurora. Omitting the last, the airglow is the strongest, being four or five times that of the other sources combined : but its detailed observation requires refined techniques using long exposure times (Chamberlain and Meinel 1954), (Armstrong and Dalgarno 1955). Both the airglow and the aurora (Hulburt 1951, Harang 1951, Swings 1949), have, however, been investigated in considerable detail by photometric and spectroscopic means and this has resulted in information on composition, temperature, and motion in the upper atmosphere.

Although most of the spectral features of the airglow have been identified, the heights of the principal sources, atomic and molecular oxygen, molecular nitrogen, sodium and hydroxyl are not known with certainty. Approximate determinations have been made by the van Rhijn method which assumes the emitting layer to be thin and uniform. Although most emissions originate at 200-300 km the height of certain of the emitting layers for atomic oxygen, e.g., 5577 Å and also that of sodium 5892 Å, appear to be rather below 100 km and that of the OH bands about 70 km. It appears, however, that accurate determination of their heights must await the results of rocket experiments using sideways-facing spectrographs. One measurement of this type has been described by Berg, Koomen, Meredith and Scolmik (1956).

Temperatures can be estimated from the spectra from measurements of Doppler widths of forbidden atomic lines, from the distribution of intensity in the rotation lines of a molecular band and by comparison of the intensities of bands with differing vibrational quantum numbers. The former is probably the most promising method since it relates directly to the kinetic temperature of the molecule. The two latter, although involving easier techniques, give the relative distributions of molecules in the rotational or vibrational levels. The kinetic temperature can only be deduced from them if Boltzmann's Law is obeyed, which is not necessarily so at the low pressures existing at the levels of emission. Recent results reported by Phillips using the first method gave 155°-231°K for the (OI) green 5577 Å line and 750°K ± 200°K for the (OI) red 6300 Å line, while Armstrong reported a temperature of about 370°K for the former (Armstrong and Dalgarno 1955). Temperatures based on intensities of rotational lines suggest about 260°K for the OH bands near 70 km, and 230°K at 100 km (Vegard 1937) and 250°-265°K at 90-140 km (Nicolet 1943) are representative values which have been obtained from observations of nitrogen bands in the aurora. In general it appears that, although interesting results are being produced by this work and further advances can be expected, the temperature measurements so far reported must be used with caution. It is considered by several authors that at present their main use is in identifying the height of the emitting layer rather than in giving a temperature at a given altitude. Another possible method of finding temperatures based on scale heights from luminosity curves and filter photographs of the aurora due to Harang (1951) has been shown on theoretical grounds by Bates and Griffing (1953) to be invalid.

Attempts have been made to measure air motion in the aurora by Meinel using photographic observations (Chapman 1954) and by the Jodrell Bank workers (Lovell 1956) using radar methods. Roach, in California, has tracked apparent motions in the airglow using isophote maps. The results of these experiments showed very high speeds (*c.* 800 m sec^{-1}) with reversals in direction before and after midnight. It appears extremely problematical whether these observed motions can be regarded as winds and their similarity with the results of radio-propagation studies of the F layer should be noted. Another reported method of obtaining winds from airglow studies is the observation of motion of the OH band around 70 km by making simultaneous photometric recordings at three stations and deducing the movement from irregular variations of the intensity patterns (Elvey 1943).

4. The temperature and zonal wind observations collected together

4.1. Temperature

Fig. 3 shows the principal temperature results collected together for the winter and summer seasons. Below about 20 km there is a good coverage over almost all latitudes from radiosonde results and the values given here should only be regarded as approximate means. The main features are that, in the troposphere, there is a rather steady lapse rate of temperature in the vertical and a decrease of temperature from equator to poles in both seasons, that in winter being considerably the greater. The high cold equatorial tropopause overlaps the low warm polar tropopause in middle latitudes. One consequence of this temperature distribution is that between 10 and 20 km the horizontal thermal gradient reverses, the poles being warmer than the equator. From 20 to 30 km the plotted values are those of Kochanski (1955). In summer the coldest region at these altitudes is at the equatorial tropopause. In winter, however, there is a region of roughly equally low temperatures *c.* 190°K at about 30 km in high latitudes. Thus there is an inversion over low and middle latitudes between 20 and 30 km in both summer and winter. At

high latitudes the inversion is found in the summer only, and in winter, from mid-latitudes to high latitudes, there is an increasing lapse rate in this altitude band. At 30 km the temperature change between seasons is only a few degrees over the equator whereas it is about 50°K over the poles. From 30 to 60 km most of the data are derived from anomalous sound-propagation experiments. They are subject to considerable errors but are consistent enough to produce the diagram shown. At about 30°N they have been supplemented by the results of rocket experiments and the searchlight measurements, and these are also sufficiently consistent to confirm the general magnitudes. Moreover, the radiosonde determinations to 40 km at about 40°N fit well into the general picture. In summer there is an inversion at all latitudes between 30 and 50 km. One surprising feature, however, is the relatively high temperature at 50-55 km at low latitudes. It would have been expected from consideration of heating by solar radiation that the maximum would be at a greater altitude over high than over low latitudes. This feature, however, is evident in the winter temperatures, the maximum value in this case occurring at about 10 km higher at the higher latitudes. In winter the inversion from 30 to 50 km occurs only at the lower latitudes and is higher, mainly from 40 to 60 km, over the poles. In spite of the lack of solar heating in the winter it appears likely that there is a region of high temperatures at 50-65 km at very high latitudes. The temperature at 50 km in low latitudes is about 280-290°K in both seasons, whereas there is probably a variation from 260°K to 300°K at higher latitudes from winter to summer in the 50-60 km altitude band. There are very few temperature data above about 60 km in either season. A mean value is given at about 30°N from the rocket experiments showing a lapse from about 55 km to 80 km with an inversion above. It will be shown below that it is possible to get more information on the probable temperature structure at these high levels from calculations based on wind data.

4.2. Wind

The wind data are more extensive than the temperature data, particularly above the region of the temperature maximum and the mean zonal components are shown in Fig. 4. Up to 20 km the information from calculations based on radiosonde temperatures and direct measurements from radar-wind finding equipment is considerable and the principal features are sketched in roughly. The main features are the strong westerly winds in the upper troposphere, those of winter being considerably stronger than those of summer. Lighter winds, often easterly, are found at both higher and lower latitudes. Detailed mean cross-sections up to 100 mb have been given by Moore (1956). Owing to the reversed horizontal temperature gradient in the lower stratosphere the westerly winds decrease considerably in intensity towards 20 km. In summer this tendency persists between 20 and 30 km at all latitudes. The effect is most marked in the sub-tropics where easterly winds are strongest in both seasons towards 30 km, because of the relatively strong horizontal thermal gradient near the cold region of the equatorial tropopause. Between 25 km and 30 km most of the wind values are from calculations from temperature soundings rather than direct measurements. Between 30 km and 60 km most of the data have been produced by the anomalous sound-propagation studies which inherently give better wind than temperature data. Other data contributing to the overall picture have been obtained from the rocket-grenade experiments, the radiosonde ascents to 40 km near 40°N and the smoke-shell experiments at 50°N. The diagram shows strong easterly winds in low latitudes in summer and very strong westerly winds at higher latitudes in winter, the maxima occurring in both cases at about 60 km. Qualitatively this distribution is consistent with the temperature fields described above. There is no great difficulty in interpreting the results of the various experiments up to 60 km or so, but at greater heights the reported

results are not always consistent and the picture shown is the best interpretation of the present evidence. The difficulty arises mainly from the fact that above 70 km the diurnal variation of wind due to tidal or thermal effects is often at least as great as the mean wind. Moreover the phase of these tidal components varies with season and height so that the derivation of a mean wind requires at least a year's observations with measurements throughout the day before the diurnal and mean winds can be separated. Most of the published observations are not sufficiently comprehensive for this to be done, and the values with a question mark are subjective estimates of the mean wind in these cases. Even when sufficient data exist there is not general agreement on how to interpret them. The reported values form a more consistent picture in summer than in winter. The movement of noctilucent clouds in high latitudes at about 80 km indicates an extension of the east winds to this level in summer, while the rocket-grenade results suggest that the easterly components are decreasing with height above about 55 km in mid-latitudes. Both the radar-meteor results and the fading records show west winds which appear to increase in strength with altitude up to at least 100 km in summer. There is some tendency for the strongest winds to be located in mid-latitudes. The sporadic E results, however, indicate easterly winds and, on the assumption that they refer to a region some 10 km higher than the fading records, there is an indication that above 110 km or so there may be a reversal in zonal direction back to mainly easterly wind components. In winter the rocket-grenade results and some of the radar-meteor measurements indicate a decrease of westerly components above about 60 km. At latitudes of 30°-40° this tendency appears to continue up to 100 km or so when easterly components appear. At 50°N however the decrease in westerly components may end at about 95 km with some reassertion of the westerly tendency above in the early winter at least, although easterlies may reappear at high levels. The general picture of easterly winds at about 50 km and westerly winds at about 100 km in summer with a reversal to westerly winds at the lower level and easterly winds at the higher level in winter in mid-latitudes would account for Johnson and Hale's (1953) finding that sound waves refracted down from the upper inversion usually were received on azimuths opposite to those from the lower inversion.

Values of both temperature and winds in spring and autumn are intermediate between the winter and summer values, the winds usually being light. There is some tendency for the temperatures in autumn around 50 km to be higher than those in spring, possibly owing to thermal-lag effects (some observations e.g., Elterman (1954) have indicated higher autumn than summer temperatures). The winds around 50 km in spring and autumn appear to be more frequently westerly than easterly, which may indicate that in the upper part of the ozonosphere the winter type of circulation, as well as being somewhat stronger, tends to persist longer than the summer type. At 30-40 km, especially at high latitudes, there are indications that the easterly flow is more persistent, especially in the spring. However, the available data at the equinoxes are fewer than at the solstices and in view of this and of the rapid changes in progress it has not been thought worth while to attempt to produce detailed cross-sections for these seasons. Above 80 km the meteor and fading records in England reveal light easterly winds at the equinoxes but stronger westerly winds are found both in summer and in winter. Although the meridional components are usually very much smaller than the zonal components at the solstices the latter are so small at the equinoxes that the wind direction then may often be mainly non-zonal.

5. A comparison of the wind and temperature data

Since enough material exists up to about 60 km to construct both wind and temperature diagrams, the question arises whether they are reasonably consistent, i.e., is the observed wind the same as that calculated from the observed temperature distribution by means of

the thermal wind equation? To test this the conditions at 25 km were regarded as established and the thermal wind equation was then applied in the form :

$$\frac{\partial T}{\partial y} = -T^2\left(\frac{2\omega \sin \phi}{g}\right)\frac{\partial}{\partial z}\left(\frac{u}{T}\right) \quad . \quad . \quad . \quad (16)$$

where $\frac{\partial T}{\partial y}$ is the horizontal temperature gradient, T the mean temperature of the layer considered, ω the earth's angular rotation, ϕ the latitude, and u the zonal wind component. For computation, 5 km steps of height were used and 10° of latitude, so that if ΔT is the horizontal temperature change over 10° latitude and u_1, T_1 refer to values at an altitude 5 km higher than the level at which u_2, T_2 are found :

$$-\Delta T = 0{\cdot}0032\, T^2 \sin \phi \left\{\frac{u_1}{T_1} - \frac{u_2}{T_2}\right\} \quad . \quad . \quad . \quad (17)$$

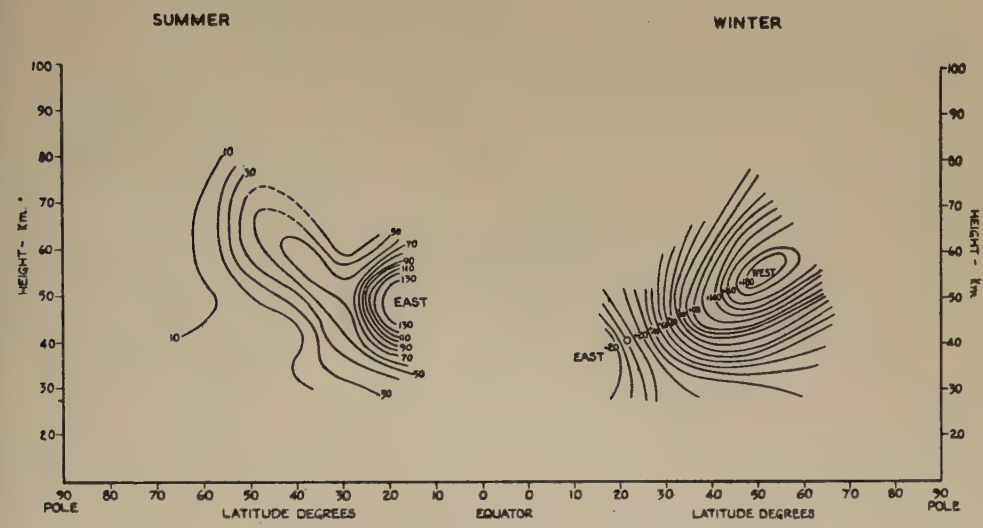

Figure 5. Winds m/sec calculated from the temperature observations.

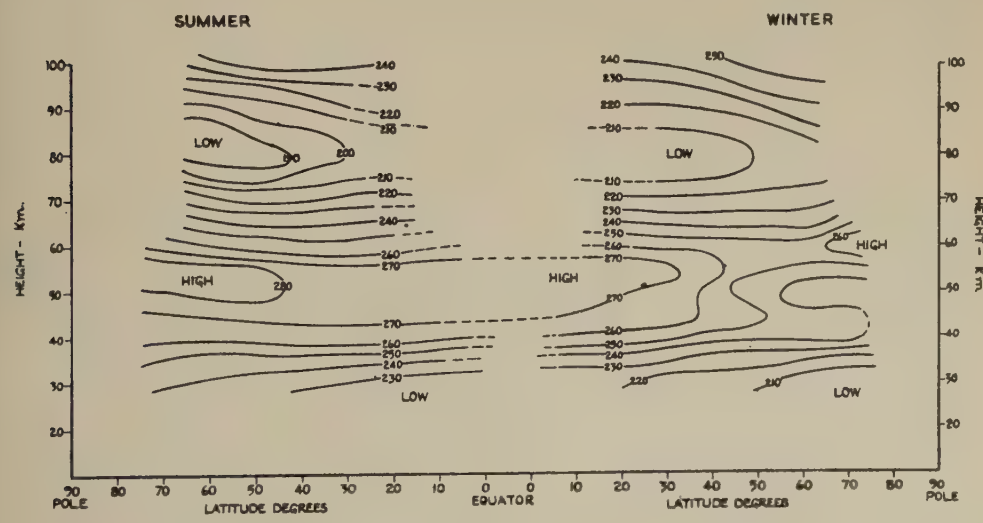

Figure 6. Temperatures °K calculated from wind observations.

Assuming the temperatures up to 60 km to be as given in Fig. 3, and the winds at 25 km to be known all the other winds were found by using Eq. (17) over 5 km steps. The results are shown in Fig. 5.

Similarly, by assuming the vertical temperature distribution at one latitude 30°N to be known and using the wind values given in Fig. 4 it was also possible by using Eq. (17) to calculate the temperature fields corresponding to the given wind distribution. The results are shown in Fig. 6. Summarizing them we find :

Summer.

Observed and calculated temperatures agree well as regards their general distribution. The main differences are lower calculated maximum temperatures at high latitudes and the non-appearance of the high-temperature region at 60 km in low latitudes in the calculated distribution.

The observed winds and those calculated from the temperature observations also agree over most of the diagram. The calculated values are excessively strong at low latitudes around 50 km. This is due partly to the failure of Eq. (16) at low latitudes and partly to the large gradients of observed temperature.

Since some wind data are available up to 100 km or so, whereas the temperature data there are confined to one latitude, it is possible to estimate the latitudinal temperature distribution above about 60 km from the wind data only. In the 60-100 km altitude band the rocket temperatures at 30° latitude were assumed to be the same in both winter and summer and the temperatures then calculated from the wind data up to 60° latitude. The results reveal the interesting feature that the lowest temperatures at 80 km are found at high latitudes in summer. The values there in winter are considerably higher, the annual range perhaps being 30°-40°C. The altitude of the lowest temperature there also appears to be higher in summer than in winter.

Winter.

The observed and calculated temperatures are in fair agreement at the region of the temperature maximum both as regards magnitudes and the increase of the altitude of the maximum at high latitudes. There is, however, an important area of disagreement at lower altitudes near the poles where the calculations from the wind data do not reveal the very low temperatures given by the direct measurements, the difference being as much as 30°C.

The comparison between the observed and calculated winds shows reasonable agreement qualitatively but very large disagreements as regards magnitudes, the computed speed being almost double that observed. The difference arises on account of the very low temperatures observed at 20 km to 40 km at high latitudes.

Wind calculations are very sensitive to the magnitude of the horizontal temperature gradients, and, therefore, in assessing which is the more representative it seems advisable to be influenced more by the wind observations than those of the temperature, since comparatively small errors in the latter produce very big effects on the former. For this reason any experimental programme aimed at producing pictures of zonal circulation would do better to make wind a main measurement rather than temperature. This can be illustrated by the table below which gives sample calculations of the vertical wind shear in 5 km of height caused by 1°C horizontal temperature differences over 10° of latitude.

In general the ratio vertical wind shear to horizontal temperature gradient is greatest at low latitudes and low temperatures, as indeed can be seen readily from Eq. (16).

TABLE 4. VERTICAL WIND SHEARS (m sec^{-1}/5 km FOR HORIZONTAL TEMPERATURE GRADIENT OF 1°C/10° lat) CALCULATED BY THE THERMAL-WIND EQUATION

Height (km)	Latitude 10°	30°	50°	70°
30	10	4	2	2
40	8	3	2	2
50	5	2·5	1·5	1·5
60	6	2·5	1·0	0·7

If we consider a 1°C error in horizontal temperature gradient extending over a height of say 30 km the effect on the wind at the top of the layer ranges from an error of 40 or 50 m sec^{-1} at 10° latitude to about 10 m sec^{-1} at 70° latitude. Thus, even at high latitudes, a small systematic temperature error is important in calculations of wind and its effect at low latitudes is predominant. The discrepancy between the observed and calculated winds at about 50 km in low latitudes does not therefore reveal any large errors in either the temperature or wind observations but merely suggests that the wind measurements should preferably be used there. The discrepancy between the winter values at high latitudes suggests, however, that there are important errors in either the adopted wind or temperature distributions. The discrepancy of almost 100 m/sec at 60 km in mid-latitudes would be reflected in an average discrepancy of 5 to 10°C in horizontal temperature over 10° latitude intervals in the 30-60 km band. Extending over 20° or 30° of latitude this accounts for the 20-30°C discrepancy between calculated and observed temperatures at these levels near the poles.

In the following section in which a standard atmosphere up to about 60 km is suggested for different latitudes in both seasons, temperature values lying between those depending on the temperature and wind observations are given, weighted roughly in the ratio of three to one in favour of the latter. This ratio is quite arbitrary but seems to give the most plausible results as regards agreement with the observational data as a whole – particularly the characteristic velocities reported in the anomalous sound-propagation experiments. The only region in which its chosen value has important effects on the resultant wind and temperature distribution is in the higher latitudes in winter. This standard atmosphere has also been extended to 100 km from 30° to 60° latitude, being based in this region almost entirely on the wind measurements.

6. A SUGGESTED STANDARD ATMOSPHERE

The adopted temperatures are listed in Table 5 and shown as a cross-section in Fig. 7. As mentioned above they were selected, mainly on considerations of horizontal temperature gradient, to provide a wind field reasonably consistent with the observations, at the same time retaining most of the features of the observed temperatures. The temperature maximum has been placed at greater altitudes in high latitudes than in low and this maximum decreases from nearly 300°K at the summer pole, through 280°K at the equator to about 260°K at the winter pole. At the mesopause the gradient appears to be in the opposite direction from about 220°K at the winter pole to 180°K at the summer pole.

The horizontal temperature gradients from about 15 to 25 km are generally negative in both seasons (temperature maximum at the poles). In winter they are large and positive around 50 km but in summer the corresponding large negative gradients occur at higher levels but do not extend over so large an area. Towards the temperature minimum around

80 km there is another reversal with large positive horizontal temperature gradients in the summer and negative in the winter. Possibly around 100 km there is yet another reversal. Nowhere does the mean seasonal horizontal temperature gradient appear to exceed about 1°C in 1° of latitude (100 km say) although doubtless this value will be considerably exceeded over shorter distances.

The dry adiabatic lapse rate is about 10°C/km and decreases only slightly with altitude. The highest lapse rates in this suggested atmosphere are at 70-80 km and do not anywhere exceed 6°C/km – which is near to the average value of the lapse rate in the troposphere. It appears doubtful, therefore, if there is widespread violent convective activity in this region. The negative lapse rates below the temperature maximum do not exceed about 4°C/km in the mean.

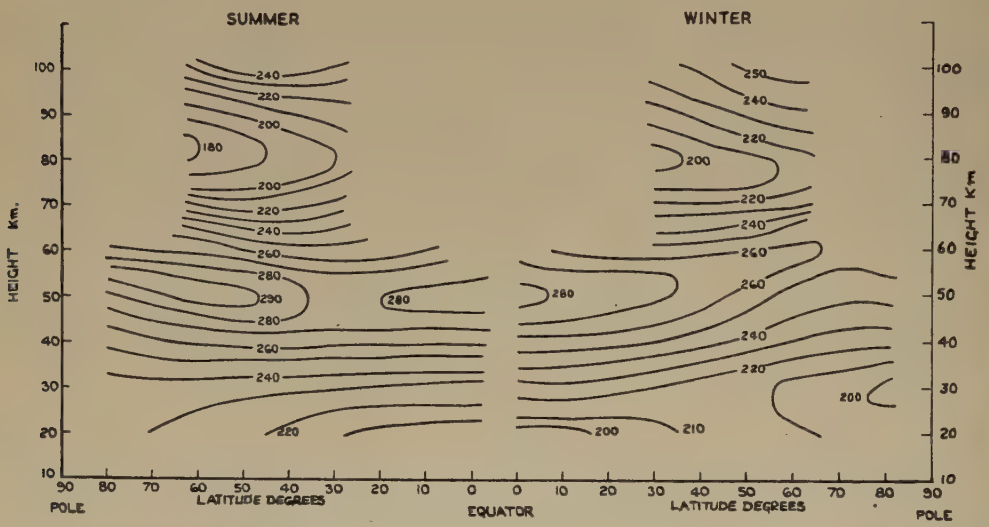

Figure 7. Adopted temperatures degrees K.

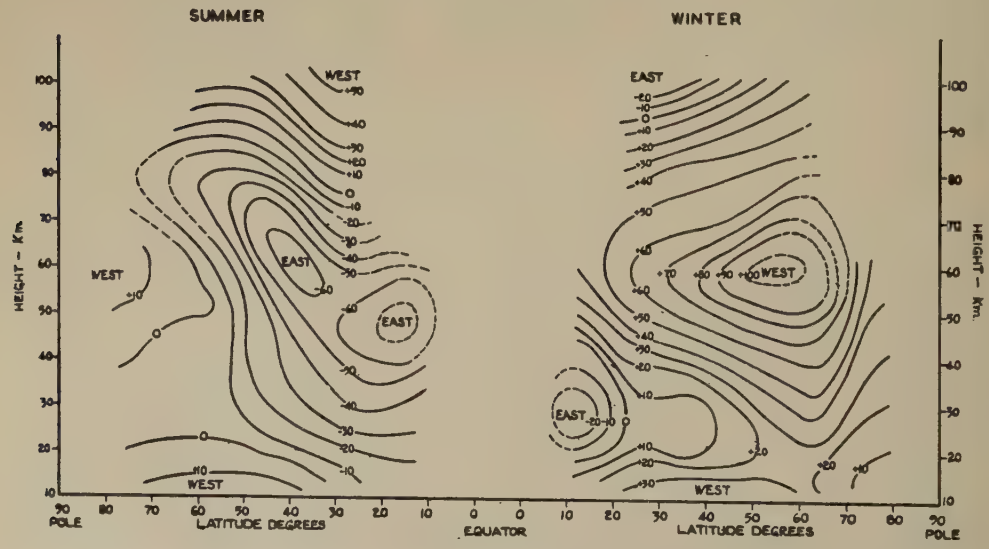

Figure 8. Zonal winds corresponding to adopted temperatures m/sec westerly positive.

TABLE 5. STANDARD ATMOSPHERE SUGGESTED TEMPERATURES (°K).

Pressure level (mb)	(a) Summer Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°	(b) Winter Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°
100	197	197	199	205	213	222	227	229	231	193	194	198	204	215	219	218	213	210
60	205	205	206	211	217	223	227	230	232	197	197	202	207	212	216	214	209	207
40	210	211	212	215	220	225	228	232	234	203	203	206	209	212	215	212	207	205
30	215	215	216	219	222	227	230	233	235	207	207	209	211	213	214	210	206	203
20	220	220	222	224	226	230	233	235	236	214	213	213	214	214	214	209	204	201
16	223	223	225	227	229	232	234	236	237	218	217	216	216	214	214	208	203	200
12	227	227	229	231	233	236	237	239	239	222	221	220	218	216	213	208	203	199
10	230	230	232	234	237	239	240	242	242	226	225	223	220	217	213	208	203	199
8	234	234	237	239	242	244	244	245	244	230	229	228	224	220	214	209	203	199
6	240	240	243	246	249	251	250	250	248	237	236	234	230	225	218	212	206	202
4	251	251	254	257	261	261	259	257	254	246	245	243	239	234	226	219	212	208
3	260	260	263	266	269	267	264	261	257	253	253	251	247	240	232	224	217	213
2·5	265	265	267	270	273	271	267	264	260	257	257	256	252	245	236	228	220	217
2·0	271	271	272	274	277	276	273	268	265	263	262	260	256	249	241	233	225	221
1·6	274	274	276	277	281	282	279	274	271	268	267	264	260	253	244	236	229	225
1·2	278	278	279	279	285	290	289	284	279	274	273	269	265	257	248	238	232	229
1·0	279	279	280	279	285	293	293	289	283	277	276	272	267	260	250	239	233	231
0·8	280	280	280	278	283	291	294	292	288	279	279	274	270	262	252	241	235	232
0·6	280	279	277	274	277	285	290	291	290	281	280	277	273	265	255	245	237	234
0·4	–	274	270	265	265	272	276	278	280	276	276	275	274	268	259	250	243	239
0·3	–	–	262	259	254	261	267	269	270	–	270	269	269	267	262	255	247	244
0·2	–	–		248	244	247	254	–	–	–	257	256	257	259	263	259	254	251
0·15	–	–	–	242	237	238	245	–	–	–		–	249	253	261	262	–	–
0·10	–	–	–	233	227	226	232	–	–		–		236	243	254	262	–	–
0·08	–	–	–	228	222	220	226	–	–		–	–	232	237	247	258	–	–
0·04	–	–	–	214	208	205	203	–	–	–	–	–	218	220	224	229	–	–
0·02	–	–	–	202	197	189	186	–	–	–	–		207	209	211	216	–	–
0·01	–	–	–	200	194	186	180	–	–	–	–	–	200	204	207	214	–	–
0·004	–	–		208	202	194	185	–	–	–	–	–	205	210	218	224	–	–
0·002	–		–	218	212	204	195	–	–	–	–	–	214	220	228	236	–	–
0·001	–	–	–	229	226	217	210	–	–	–	–	–	224	233	239	244	–	–
0·0005	–	–	–	241	243	238	234	–	–	–	–	–	236	245	250	250	–	–

TABLE 6. HEIGHTS OF PRESSURE CONTOURS (geodynamic km)

Pressure level (mb)	(a) Summer Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°	(b) Winter Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°
100	16·6	16·6	16·7	16·8	16·8	16·7	16·6	16·5	16·5	16·6	16·6	16·5	16·4	16·2	15·9	15·7	15·4	15·2
60	19·6	19·6	19·7	19·9	20·0	20·0	20·0	20·0	20·0	19·5	19·5	19·5	19·5	19·4	19·2	18·9	18·5	18·4
40	22·1	22·1	22·2	22·4	22·6	22·7	22·7	22·7	22·7	21·9	21·9	22·0	22·0	21·9	21·7	21·4	21·0	20·8
30	23·9	23·9	24·0	24·3	24·5	24·6	24·6	24·7	24·7	23·6	23·6	23·7	23·7	23·7	23·6	23·2	22·8	22·6
20	26·5	26·5	26·6	26·9	27·1	27·3	27·4	27·5	27·6	26·1	26·1	26·2	26·2	26·2	26·1	25·7	25·2	25·0
16	27·9	28·0	28·1	28·4	28·6	28·8	28·9	29·0	29·1	27·5	27·5	27·6	27·6	27·6	27·5	27·1	26·6	26·3
12	29·8	29·9	30·0	30·3	30·6	30·8	30·9	31·0	31·1	29·4	29·4	29·5	29·5	29·4	29·3	28·8	28·3	28·0
10	31·1	31·1	31·2	31·6	31·8	32·1	32·2	32·3	32·4	30·6	30·6	30·7	30·6	30·6	30·4	30·0	29·4	29·0
8	32·6	32·6	32·8	33·1	33·4	33·7	33·8	33·9	34·0	32·1	32·1	32·2	32·1	32·0	31·8	31·3	30·7	30·3
6	34·6	34·6	34·8	35·2	35·5	35·7	35·9	36·0	36·1	34·1	34·0	34·1	34·0	33·9	33·7	33·1	32·4	32·0
4	37·5	37·5	37·8	38·2	38·5	38·8	38·9	39·0	39·1	37·0	37·0	37·0	36·8	36·6	36·3	35·7	34·9	34·5
3	39·7	39·7	40·0	40·4	40·7	41·0	41·1	41·2	41·2	39·1	39·0	39·0	38·9	38·7	38·2	37·6	36·7	36·2
2·5	41·1	41·1	41·4	41·8	42·2	42·5	42·5	42·6	42·6	40·4	40·4	40·4	40·2	40·0	39·5	38·8	37·9	37·6
2·0	42·8	42·9	43·2	43·6	44·0	44·2	44·3	44·4	44·3	42·1	42·1	42·1	41·9	41·6	41·1	40·3	39·3	38·8
1.6	44·6	44·6	44·9	45·4	45·8	46·1	46·1	46·1	46·0	43·9	43·8	43·8	43·6	43·2	42·6	41·8	40·8	40·3
1·2	47·0	47·0	47·3	47·7	48·2	48·5	48·5	48·5	48·4	46·2	46·1	46·1	45·8	45·4	44·7	43·8	42·8	42·2
1·0	48·4	48·5	48·8	49·2	49·7	50·1	50·1	50·0	49·9	47·6	47·6	47·5	47·2	46·8	46·1	45·1	44·0	43·5
0·8	50·3	50·3	50·6	51·1	51·6	52·0	52·0	51·9	51·8	49·5	49·4	49·3	49·0	48·5	47·7	46·7	45·6	45·0
0·6	52·7	52·7	53·0	53·4	54·0	54·4	54·5	54·4	54·2	51·8	51·8	51·6	51·3	50·7	49·8	48·7	47·6	47·0
0·4	–	56·0	56·2	56·6	57·2	57·7	57·8	57·8	57·6	55·2	55·1	55·0	54·5	53·9	52·9	51·7	50·4	49·7
0·3	–	–	58·5	58·8	59·4	60·0	60·1	60·1	59·9	–	57·4	57·2	56·8	56·1	55·1	53·8	52·5	51·8
0·2	–	–	–	61·8	62·3	63·0	63·2	–	–	–	60·5	60·4	60·0	59·3	58·3	56·9	55·5	54·8
0·15	–	–	–	63·9	64·4	65·1	65·4	–	–	–	–	–	62·1	61·4	60·4	59·1	–	
0·10	–	–	–	66·7	67·1	67·8	68·2	–	–	–	–	–	65·0	64·4	63·6	62·2		–
0·08	–	–	–	68·3	68·6	69·3	70·0	–	–	–	–	–	66·5	66·0	65·2	63·9	–	–
0·04	–	–	–	72·7	73·0	73·6	74·0	–	–	–	–	–	71·1	70·6	70·0	68·9		–
0·02	–	–	–	77·0	77·1	77·6	78·0	–	–	–	–	–	75·4	75·0	74·4	73·5		
0·01	–	–	–	81·1	81·1	81·4	81·8	–	–	–	–	–	79·6	79·2	78·6	77·8	–	–
0·004	–	–	–	86·6	86·4	86·5	86·7	–	–	–	–	–	85·0	84·7	84·3	83·7	–	
0·002	–	–	–	90·9	90·6	90·6	90·5	–	–	–	–	–	89·3	89·1	88·9	88·4	–	–
0·001	–	–	–	95·4	95 1	94·9	94·7	–	–	–	–	–	93·7	93·7	93·6	93·3		–
0·0005	–	–	–	100·2	99·8	99·5	99·2	–	–	–	–	–	98·4	98·6	98·6	98·3	–	–

TABLE 7. STANDARD ATMOSPHERE DENSITIES, ($\log_{10} \rho$ g cm^{-3} negative values)

Pressure level (mb)	(a) Summer Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°	(b) Winter Latitude 0°	10°	20°	30°	40°	50°	60°	70°	80°
100	3·75	3·75	3·76	3·77	3·79	3·80	3·89	3·82	3·82	3·74	3·75	3·76	3·77	3·79	3·80	3.80	3·79	3·78
60	3·99	3·99	4·00	4·00	4·02	4·03	4·04	4·04	4·05	3·97	3·97	3·99	4·00	4·01	4·01	4·01	4·00	4·00
40	4·18	4·18	4·18	4·19	4·20	4·21	4·21	4·22	4·23	4·16	4·16	4·17	4·18	4·18	4·19	4·18	4·17	4·17
30	4·31	4·31	4·32	4·32	4·33	4·34	4·34	4·35	4·35	4·30	4·30	4·30	4·31	4·31	4·31	4·30	4·29	4·29
20	4·50	4·50	4·50	4·51	4·51	4·52	4·52	4·53	4·53	4·49	4·49	4·49	4·49	4·49	4·49	4·48	4·47	4·46
16	4·60	4·60	4·61	4·61	4·61	4·62	4·62	4·63	4·63	4·59	4·59	4·59	4·59	4·49	4·58	4·57	4·56	4·56
12	4·74	4·74	4·74	4·74	4·75	4·75	4·75	4·76	4·76	4·73	4·72	4·72	4·72	4·71	4·71	4·70	4·69	4·68
10	4·82	4·82	4·82	4·83	4·83	4·83	4·84	4·84	4·84	4·81	4·81	4·81	4·80	4·80	4·79	4·78	4·77	4·76
8	4·92	4·92	4·93	4·94	4·94	4·94	4·94	4·95	4·94	4·92	4·92	4·91	4·91	4·90	4·89	4.88	4·88	4·85
6	5·06	5·06	5·07	5·07	5·08	5·08	5·08	5·08	5·07	5·05	5·05	5·05	5·04	5·03	5·02	5·01	5·00	4·99
4	5·26	5·26	5·26	5·27	5·27	5·27	5·27	5·27	5·26	5·25	5·25	5·24	5·23	5·23	5·21	5·20	5·18	5·17
3	5·40	5·40	5·40	5·41	5·41	5·41	5·40	5·40	5·39	5·38	5·38	5·38	5·37	5·36	5·35	5·33	5·32	5·31
2·5	5·48	5·48	5·49	5·49	5·50	5·49	5·49	5·48	5·48	5·47	5·47	5·47	5·46	5·45	5·43	5·42	5·40	5·40
2·0	5·59	5·59	5·59	5·60	5·60	5·60	5·59	5·59	5·58	5·58	5·58	5·57	5·57	5·55	5·54	5·52	5·51	5·50
1·6	5·69	5·69	5·70	5·70	5·70	5·70	5·70	5·69	5·69	5·68	5·68	5·68	5.67	5·66	5·64	5·63	5·61	5·61
1·2	5·82	5·82	5·82	5·82	5·83	5·84	5·84	5·83	5·82	5·82	5·82	5·81	5·80	5·79	5·77	5·76	5·74	5·74
1·0	5·90	5·90	5·91	5·90	5·91	5·93	5·93	5·92	5·91	5·90	5·90	5·89	5·89	5·87	5·86	5.84	5·83	5·82
0·8	6·00	6·00	6·00	6·00	6·01	6·02	6·02	6·02	6·01	6·00	6·00	6·00	5·99	5·97	5·96	5·94	5·93	5·92
0·6	6·13	6·13	6·12	6·12	6·12	6·13	6·14	6·14	6·14	6·13	6·13	6·12	6·12	6.10	6·09	6·07	6·05	6·05
0·4	–	6·29	6·29	6·28	6·28	6·29	6·30	6·30	6·30	6·30	6·30	6·30	6·29	6·28	6·27	6·25	6·24	6·23
0·3	–	–	6·40	6·39	6·39	6 40	6·41	6·41	6·41	–	6·41	6·41	6·41	6·41	6·40	6·39	6.37	6·37
0·2	–		–	6·55	6·54	6·55	6·56	–	–	–	6·57	6·57	6·57	6·57	6·57	6·57	6·56	6·56
0·15	–	–	–	6·67	6·66	6·66	6·67	–	–	–	–	–	6·68	6·69	6·70	6·70	–	–
0·10		–		6·83	6·81	6·81	6·82	–	–	–	–	–	6·83	6·84	6·86	6·88		
0·08	–		–	6·91	6·90	6·90	6·91	–	–	–	–	–	6·92	6·93	6·95	6·97		
0·04	–	–		7·19	7·17	7·17	7·16	–	–	–	–	–	7·20	7·20	7·21	7·22		
0·02	–	–	–	7·46	7·45	7·43	7·43	–	–	–	–	–	7·47	7·48	7·48	7·49		–
0·01	–	–		7·76	7·75	7·73	7·71	–	–	–	–	–	7·76	7·77	7·77	7·79		
0·004	–		–	8·17	8·16	8·14	8·12	–	–	–	–	–	8·17	8·18	8·19	8·21		
0·002		–		8·50	8·48	8·47	8·45	–	–	–	–	–	8·49	8·50	8·52	8·53	–	
0·001	–	–	–	8·82	8·81	8·80	8·78	–	–	–	–	–	8·81	8·83	8.84	8·85	–	–
0·0005	–		–	9·14	9·14	9·14	9·13	–	–	–	–	–	9·13	9·15	9·16	9·16	–	–

The annual temperature ranges shown in Fig. 9 reveal very little variation between summer and winter in low latitudes. At high latitudes there is a variation of about 30°C at most altitudes, and a maximum around 40°C near 50 km. At 80 or 90 km the surprising result has been obtained that the annual variation is in the opposite direction, winter temperatures being considerably higher than summer temperatures. This, of course, follows from the assumption that there is little annual variation at low latitudes and the reversal of the horizontal temperature gradient.

The zonal winds calculated from the adopted temperatures are shown in Fig. 8. It contains the major features of the diagram of observed winds although some of the irregularities in the former are smoothed out. The two major disagreements are the rather high values of the east wind at about 45 km in very low latitudes in summer and the very high values of westerly winds in mid-latitudes at about 60 km in winter. The former can be accounted for to some extent by the failure of the geostrophic-wind equation at low latitudes. The latter, however, although it has not been directly observed seems to be very plausible from the observed temperatures. It has some indirect support from the results of the rocket-grenade experiments in lower latitudes (Weisner 1956) where winds of 90-100 m sec^{-1} were found in some experiments and also from the occasional very high value of the characteristic velocity of sound waves reported in anomalous sound-propagation experiments (Murgatroyd 1956a).

For many purposes it is desirable to relate temperatures, winds and densities to pressure levels. To the accuracy of these data the height of the pressure levels can be expressed either in geometric (z) or geodynamic (h) units. If E is the radius of the earth then

$$h = \frac{g_0}{10} \frac{z}{1 + (z/E)} \quad . \quad . \quad . \quad . \quad (18)$$

where the gravitational acceleration g_0 can be taken as equal numerically to 9·8 m/sec^2. By using h instead of z in integrating the hydrostatic equation (4) the variable gravitational acceleration g is eliminated. The integrations were made upwards from the 100 mb level where mean values of geodynamic heights and temperatures at the different latitudes were supplied by the Upper Air Climatology Branch of the Meteorological Office. The successive pressure surfaces upwards were chosen to give height intervals of 2-3 km up to about 60 km and 3-5 km above, mean temperatures within each layer being used. Resultant contour heights are listed in Table 6.

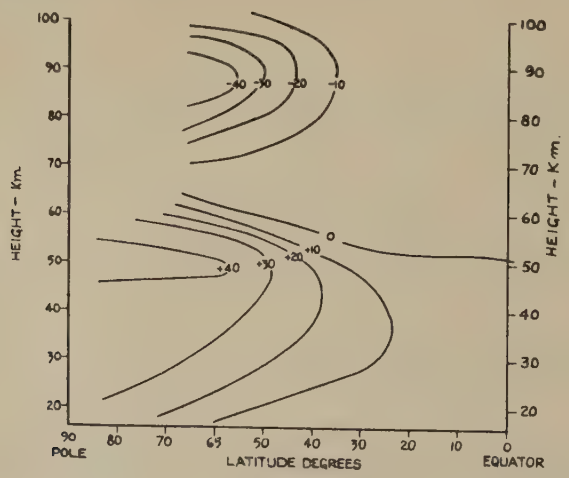

Figure 9. Annual temperature range, degrees K (Summer minus winter temperatures).

Densities were found from the pressure and temperature at each level assuming constant composition of the air throughout. Values are given in Table 7.

Finally, the zonal wind components shown in Fig. 8 were found at intermediate latitudes e.g. 15°, 25°, 35°, etc., by using the geostrophic-wind equation at 10° latitude intervals in the form :

$$u = \frac{10}{2\omega \sin \phi} \frac{dD}{dy} \quad . \quad . \quad . \quad . \quad (19)$$

where $\frac{dD}{dy}$ is the slope of the isobaric surface.

7. Concluding remarks

The above discussion has been confined mainly to the distribution of temperature and zonal wind components. There are many other features which should be dealt with, e.g., meridional components, size of pressure systems, field of vertical motion and degree of turbulence.

Unfortunately the necessary data do not yet exist in many cases and only the most general conclusions can, as yet, be drawn. Kellogg, and Schilling (1951), Goldie (1950) and Palmer (1954) have attempted to deduce meridional circulations but their results at this stage must be regarded as extremely tentative. Similarly, although it is possible to obtain preliminary ideas on the degree of turbulence by use of a large-scale Richardson Number (Sheppard 1954) we need to know the wind structure in considerably greater detail before much progress can be made. The meteorologist dealing with the atmosphere above the tropopause to 100 km or so must consider conditions in four distinct regions, i.e., the lower stratosphere, say below 30 km, the region where the temperature increases with height (30-60 km), the altitude band where the temperature decreases with height (60-80 km), and the second inversion layer above (80 km upwards). The principal features of these regions may be summarized :

(i) The lower stratosphere has some coupling to the troposphere in that ozone amounts, temperature and winds can be correlated with tropopause conditions, but the coupling decreases with altitude (Bannon and Jones 1955). Wind systems in this region are similar to those of lower levels with sizeable meridional as well as zonal components, small-scale variations (Kochanski and Wasko 1956) and considerable longitudinal variation (Bannon and Gilchrist 1956).

(ii) The inversion region from 30-60 km appears to have very little coupling with the troposphere. Friction effects and turbulence are probably small, tidal effects are inconsiderable and pressure, temperature, and wind relationships are similar to those at lower levels. Familiar atmospheric phenomena such as long waves in the general flow and possibly persistent jet streams may be expected but as yet there appears to be little evidence of smaller systems on the scale of the cyclones and anticyclones found at lower levels. Diurnal changes of temperature of a few degrees are likely depending on the solar irradiation. There is evidence also of short-period wind variations. No systematic variation of meridional components of wind has emerged from the various measurements, and the zonal components are usually stronger and more persistent in character.

(iii) From 60 to 80 km the lapse rate appears to be large enough for convection to take place, particularly over high latitudes in summer, although it is doubtful whether the dry adiabatic lapse rate is exceeded. The evidence that the mesopause is at its coldest above the hottest part of the ozonosphere parallels conditions at the tropopause which is also at its coldest (and highest) above the hottest region below. Pant (1956) has recently attempted to explain this qualitatively in terms of the infra-red radiation balance following

the lines of the discussion of Dobson, Brewer and Cwilong (1946) of the radiation temperature of the tropopause. The winds in this region have regular seasonal features with the zonal components larger and more steady than the meridional component. Tidal components start to be easily measurable at the top of this region, their importance increasing with altitude. Little can be inferred about the size of the pressure systems here but considerable turbulence and large wind shears are a notable feature of all reported observations.

(iv) Turbulence is also found in regions from 80 km upwards although the lapse rate is very stable there. The evidence for this arises mainly from observations of meteors and it is not altogether clear to what extent these merely reflect the meteor's disturbance of the atmosphere or whether they are representative of atmospheric conditions. This could be tested by observing whether the turbulence measured decreases with time in given portions of long-enduring trails. The problem of wind measurement in this altitude band is complicated by the large oscillatory components but the radar-meteor technique of wind finding has made it possible to extract the prevailing winds. Zonal components have again been shown to be predominant although their seasonal changes of direction require explanation. The variation of the oscillatory components with altitude and season outlined in § 3.8, although complex, is in reasonable agreement with the predictions of the resonance theory.

As regards the future, the first requirement is for more observational data at all levels and it appears that a good deal will soon be forthcoming, particularly from increased high-level balloon ascents, rockets, and the radio-meteor and radio-propagation studies. Similarly, further measurements of atmospheric composition, particularly ozone and water vapour at all possible levels, are required to allow studies of the radiation field to be further advanced. As these become available, rapid progress in our understanding of the circulation of the upper atmosphere can be expected in the next few years.

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Further studies in World climate

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SUMMARY

The effects of changes in solar radiation on the climate of the earth are discussed with special reference to glacial phases in high latitudes and to pluvial periods which are common to the Earth as a whole. The relationship between the radiation and the glacial and pluvial phases is expressed in tabular form. Adopting the chronology of the glacial and pluvial phases during the Pleistocene Period, as published by K. P. Oakley, a curve of solar radiation is constructed using only the glacial phases; this proves to be a periodic curve with three complete oscillations between the begining of the Pleistocene Period and the present day. Using this radiation curve the periods of heavy and light precipitation are determined and compared with the pluvial periods. Three periods of heavy precipitation during the Pleistocene are indicated and a correlation of these with the observed pluvial periods in Africa is suggested.

1. INTRODUCTION

In 1934 I read before the Society a paper entitled 'World climate during the Quaternary Period' in which I set out and discussed in considerable detail a theory of climate changes based on postulated variations in solar radiation. In the twenty years since then the theory has been much discussed without any general acceptance or rejection.

When I wrote that paper it was generally accepted that there had been four glacial periods (now called 'phases'): Günz, Mindel, Riss and Würm in high northern latitudes and two pluvial periods in low latitudes during the Pleistocene Period. There was much doubt whether the glacial phases were simultaneous in the north and south hemispheres, or even in North America and Europe; how many pluvial periods there had been during the Pleistocene and whether they coincided with the glacial or interglacial phases.

Since then a great deal of work has been done by geologists and prehistorians on the climate of the Pleistocene in most parts of the world and especially in Africa. Some of the problems have been solved: it is now generally accepted that glacial phases were simultaneous in all parts of the Earth, and four pluvial periods have been distinguished in East Africa. On the other hand new difficulties have arisen: there is now doubt about the four glacial phases, signs of further glacial phases before the Günz having been found, and only three pluvial phases can be recognized in large parts of Africa. The workers in East Africa seem agreed that the four pluvial phases recognized by them – Kageran, Kamasian, Kanjeran and Gamblian – each coincided in time with the four glacial phases of Europe and America.

In 1954 a 'Pelican' book on '*The prehistory of East Africa*' by Sonia Cole, who herself had worked with Dr. Leakey in East Africa, was published which gave an excellent account of the recent work carried out in East Africa and made it possible to compare the prehistory of Africa with that of the rest of the world. Reading this book reawakened my interest in a problem to which I had given little thought for nearly twenty years. Considering that two of the results of my earlier work were not confirmed – four instead of my two pluvial phases and the coincidence of the pluvial phases in Africa with the glacial phases in Europe – I naturally reviewed my previous work and studied the implications of the new evidence only to be more convinced than ever that the solar radiation theory is built on sound foundations. It is the results of this further study that I wish to report in this paper.

In my 1934 paper I first suggested a cause of an ice-age, then built up around it a purely theoretical scheme which I hoped to prove by describing the geological evidence which I considered supported it. This proved an unfortunate way, for more criticism was devoted to my geology than to my meteorology. In this paper I shall make more use of the observations in building up and testing my argument.

The plan I intend to follow is to give a short outline of the solar radiation theory and then use it in a discussion of the effects of changes in the radiation on the climate under the headings Temperature, Precipitation, Glacial Phases and Pluvial Phases. The results of the discussion will be summarized in a diagram, Fig. 1, showing the relationship between the sequences of the glacial and pluvial phases and their dependence on the changes in solar radiation (the ' solar constant ').

The best available results of the work of geologists and prehistorians on the chronology of the glacial and pluvial phases are then discussed. Having assembled all the necessary data, those relating to the sequence of glacial phases alone are then used to construct a curve of radiation; this curve is in turn used to determine a sequence of pluvial phases for comparison with the geological observations in Africa. Thus starting with geological evidence of glacial phases in high latitudes the theory is used to obtain a sequence of pluvial phases for comparison with the geological evidence of pluvial rainfall in low latitudes. The results of the investigation are discussed in the last chapter.

2. The solar radiation theory

The temperature of the Earth as a whole is controlled by the radiation received from the sun and is determined by the temperature at which the total outgoing long-wave terrestrial radiation equals the total incoming short-wave solar radiation. Of the total amount of solar radiation reaching the earth a certain proportion, called the albedo, is reflected as short-wave radiation, mainly from the clouds, and therefore does not add any heat; the remaining solar radiation enters the atmosphere and is absorbed either by the atmosphere or the surface. This portion is called the ' effective solar radiation.'

The outgoing ' terrestrial radiation ' originates on the Earth's surface, on the water vapour in the atmosphere and on the water particles in the clouds; all governed by the temperature where the radiation originates.

As the major changes in solar radiation are extremely slow – we are discussing change in which the unit of time is a thousand years – the Earth as a whole is always in radiative balance, the existing temperatures of the surface, atmosphere and clouds constantly adjust themselves, mainly through the action of the general circulation of the atmosphere, until the balance is reached and maintained.

How the adjustment is reached according to the theory is best considered by studying what changes would automatically take place if the solar radiation were suddenly increased by a definite amount. The amount of incoming solar radiation is specified by the intensity of the radiation at the confines of the atmosphere and is measured by the energy of the radiation which falls on a square centimeter of surface exposed at right angles to the incoming radiation just outside the confines of the atmosphere; it is expressed in calories per square centimeter per minute and is called the ' solar constant.' At present the solar constant is $1{\cdot}95$ cal cm^{-2} min^{-1}.

The first result of an increase in the solar constant would be a general increase in the effective radiation received in all parts of the Earth; this, acting on the existing conditions, would raise the temperature of the surface and of the atmosphere. This increase would be greater in equatorial than in polar regions, so that the temperature gradient between the equator and the pole would increase, producing a more active general circulation of the

atmosphere. The result of these changes – a general increase in temperature and an increase in total wind – would be an increase in evaporation from both land and water. But higher surface temperatures, more wind and more evaporation can only result in more cloud and more precipitation, for all the water evaporated must return to the surface in the form of either rain or snow.

These changes in the atmosphere, however, have a reaction on the radiation both incoming and outgoing. In the first place the larger amount of cloud increases the albedo, so that a larger proportion of the incoming radiation is reflected, there is some increase in effective radiation, but it is not so large as the increase in the solar constant; this increase in the effective radiation raises the temperature of the surface and of the atmosphere and increases the outgoing terrestrial radiation. Acting together all these factors – increased solar constant, increase in the temperature of surface and atmosphere and increase in cloud amount – produce a new balance of radiation which will be maintained until there is another change in solar constant. These are the primary changes due to an increase in solar radiation and there is nothing remarkable about them; they are the natural physical effects which one would expect; it is in the secondary effects of these primary changes that the paradox of the solar radiation theory lies. It is that the increase of solar radiation leads to the glaciation of the so-called Ice Ages. This, however, is not difficult to explain.

The increase of cloud which follows the increase in solar radiation will occur chiefly in the three main cloud belts, one over equatorial regions and two in high north and south latitudes respectively. In the latter, the chief rise in temperature would occur in the winter, caused not by increased radiation (of which there is very little during the winter in high latitudes), but by greater inflow of warmer air from lower latitudes bringing with it an increase of water vapour to give increased snowfall. On the other hand the increased cloud amount reduces the amount of solar radiation reaching the surface in the summer, so reducing the amount of melting.

It has constantly been pointed out in the literature that (a) a decrease of temperature alone would not produce an ice-age; for it would reduce the precipitation while all the evidence, especially in the Antarctic, is that a considerable increase in the present snowfall would be necessary to build up the thick sheets of ice which covered the polar regions in a glacial phase, and that (b) as the extent of an ice-sheet is controlled more by the melting accompanying direct sunshine in the summer than by the temperature of the air, a reduction of the hours of sunshine by an increase of cloud amount would be an important factor in the maintenance of an ice-sheet once formed. Thus provided that the temperature of the polar regions remained below the freezing point for a considerable part of the year, the conditions which would accompany an increase in solar radiation – more snow in the winter and less sunshine in the summer – are just those which are required to produce glaciation in high latitudes and on high mountains.

A nucleus ice-cap having formed at the Pole in this way it will grow as the radiation increases owing to the ever increasing snowfall on the ice-cap itself. But as the margin of the cap extends it reaches into warmer latitudes and increased summer melting is met with. Without going into details the remainder of the story is soon told. At first the ice-cap expands because the increase in snowfall is more important than the increased melting due to increased radiation, reduced as the latter is by increased cloud cover; but there comes a time when these effects become equal after which the increased melting at the margin due to the ever increasing radiation will take the upper hand and the margin will retreat until the ice-cap becomes very small and finally disappears entirely.

If the radiation now commences to decrease, the whole process is gone through again in the reversed order. There will probably be some hysteresis, but that must be disregarded, if for no other reason than that we know nothing about it and the problem is too complex

to be solved by qualitative reasoning. The radiation theory therefore *assumes* that each of the climates described above occurs when the solar constant has a certain value and it is immaterial whether the radiation approached that value by increasing or decreasing. In all this it is understood that there have been no geographical changes sufficiently large to have their own effects; thus the theory should not be used in its simple form to solve questions of climate in all periods of the geological past. In this paper it is restricted to the Pleistocene Period during which we know that there were no such large geographical changes.

We have now to apply this theory to the actual conditions on the Earth and discuss the relationship between the solar radiation and the three main factors of climate : temperature, precipitation and glaciation. The most convenient way to do this will be to prepare a diagram setting out the results and to explain it as the discussion proceeds. Fig. 1 has been prepared for this purpose.

The diagram is divided into six main columns lettered (a) to (f). In the first column (a) there is a scale, not necessarily linear, of solar radiation expressed as solar constant. As we only know the numerical value of the solar constant for the present time, 1·95 cal cm^{-2} min^{-1}, this value has been entered at a convenient point on the scale S_B. Other points on the scale referred to are labelled S_C, S_D, etc., in order of increasing magnitude. The other columns will be described as we use them in the course of the discussion.

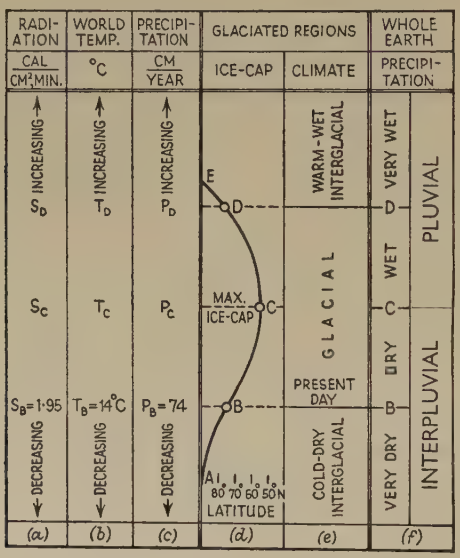

Figure 1. Radiation and climates.

3. Temperature

When the present mean annual temperature at sea-level of each circle of latitude (the zonal temperature) is plotted against the latitude the result is a smooth curve increasing from the Pole to the Equator in each hemisphere (see Fig. 5 (a) for the zonal temperatures of the northern hemisphere). The remarkable fact is that in spite of the great differences between the amounts of land and sea in the two hemispheres and along different circles of latitude, the two curves for the northern and southern hemispheres are practically identical; each is a smooth curve, the difference between them never being greater than

3°C at any latitude. Judging from the observations nothing on the earth's surface, land or water, forest or grass, desert or swamp, affects the mean temperature of a zone sufficiently to break the smooth run of the curve. In fact the present distribution of the zonal temperature appears to be independent of the nature of the Earth's surface (Simpson 1927).

This does not mean that the temperature at every point along a circle of latitude is the same; the observations show that there are large variations in temperature along a circle of latitude (other than those due to height above sea-level which are removed by reducing all observed temperatures to sea-level). The difference of the temperature at a place from the mean annual temperature of the circle of latitude on which it is situated is called the thermal anomaly of that place and charts of the world have been prepared in which the distribution of the thermal anomalies is shown by isanomalies (Hann – Suring 1926, p. 151). These charts show large areas in which all the departures are positive and others in which they are all negative. A study of these isanomalies shows that the anomalies are generally due to ocean currents bringing cold or warm water which lowers or raises the temperature locally, to deserts and especially ice-sheets affecting the radiation balance.

Now the decrease of temperature from equator to pole is a consequence of the shape of the earth governing the distribution of the incoming solar radiation. This temperature gradient between equator and pole sets up the general circulation of the atmosphere which transfers heat from equatorial to polar regions and smooths out the thermal anomalies until a radiation balance is obtained. Thus the existing thermal distribution with its zonal temperatures is the result of solar radiation acting on a spherical earth, almost unaffected by the nature of the surface.

There is no reason to doubt that at all times since the Earth had an atmosphere and the sun was the only source of heat the temperature distribution was similar to the present; the mean temperature of the Earth as a whole may have been higher or lower according to the amount of solar radiation received from the sun; and the temperature gradient between the equator and the poles may have been greater or less; but the general shape of the curve of zonal temperature would always have been similar; that is the zonal temperatures have always decreased from a maximum in the equatorial zone to a minimum in the polar zone, every zone being colder than the zone on its equatorial side and warmer than the zone on its polar side; and this must have been the case whether the surface was mainly land, mainly water or largely ice.

To sum up : although we do not know how much the temperature of any one place or the mean temperature of any one latitude will be affected by a change in the solar constant, we do know that there must be a change in the same direction in the *mean temperature of the world as a whole*, or there could be no radiation balance, and that the zonal temperature will decrease from equator to pole in a smooth curve similar to the one at the present time. We shall make much use of these relationships. We have now to enter this information in Fig. 1.

As each point on the ordinate for radiation in (a) of Fig. 1 represents a definite value of the solar constant, the values increasing upwards, and as every value of the solar constant has a definite value of the mean temperature of the earth associated with it, the scale of temperature in (b) must also increase continuously in the upward direction, but not necessarily linearly. At the present time the solar constant is 1·95 cal cm^{-2} min^{-1} and has already been entered in (a); the mean temperature of the earth as a whole (taking account of the area of each zone) is 14°C and this value is entered in (b). Other values of the mean temperature of the Earth in degrees centigrade are written T with subscripts, T_A, T_B etc. to specify the temperature represented.

If we had a few more numerical values of the mean temperature at epochs with materially different solar constants we might be able to deduce the form of the relationship between them and fill columns (a) and (b) with numerical values. Meanwhile the fact that the mean temperature of the Earth rises and falls with the solar constant is all we know about the relationship between solar radiation and terrestrial temperature; but this will be sufficient for our present purpose.

4. Precipitation

Over a long period of time the average annual amount of precipitation which falls on the whole surface of the Earth must be equal to the average annual total of water evaporated. The amount of water evaporated from the Earth's surface depends on two main factors : (a) the temperature of the air at the surface and the general circulation of the atmosphere. Now both of these factors are governed by solar radiation, the temperature directly and the circulation through the temperature gradient between the equator and the poles. The total evaporation must therefore increase and decrease as the solar radiation increases and decreases; and with it the total precipitation also.

The distribution of the precipitation over the Earth is a consequence of the general circulation of the atmosphere. We know nothing directly as to how this distribution would be affected by variations in the circulation caused by changes in the radiation; but there is no reason to believe that the distribution would vary in essentials. With an increase in radiation precipitation would increase everywhere where precipitation is now appreciable and spread into regions which are now dry. We may conclude from this that the total amount of precipitation over the Earth as a whole, like temperature, increases and decreases with the solar constant; but this does not apply to individual zones or areas especially where the climate is affected by the proximity of ice-sheets; in particular local variations of precipitation have little or no relation to the changes in radiation.

The sources of information about past changes in the amount of precipitation are mainly geological and are not capable of giving numerical values of the precipitation; hence we can only describe the amount of precipitation in words or phrases, generally comparisons with the precipitation of the present time. We must again be satisfied with entering in column (c) of Fig. 1, which represents a scale of precipitation, the present value, 74 cm/year (Hann-Suring 1926, p. 373) and indicate that the total precipitation increases and decreases with the radiation. Individual amounts on the scale are represented by the letter P with the appropriate subscript.

5. Glacial and interglacial phases

In § 2 on the solar radiation theory I have given a short description of how a polar ice-cap forms, grows and then melts away as the solar radiation increases from low to high values. We have now to examine the present north polar ice-cap and the evidence we have of previous ice-caps in the light of this theory.

Geologists recognize four main glacial phases during the Pleistocene Period, in each of which an ice-cap formed in polar regions and mountains in all parts of the world were heavily glaciated with glaciers descending far below the present snow-line. These glacial phases were first recognized in the Alps by Penck and Brückner (1909) and were named by them in order of age Günz, Mindel, Riss and Würm.

The map Fig. 6 shows the extent of the north polar ice-cap at present and at the maximum of the Riss glaciation. Owing to the fact that the North Pole is situated in the middle of a deep ocean, a solid permanent ice-sheet cannot form. But in the central parts

of the Arctic Ocean the ice is so closely packed and so thick that it is not possible for ships to penetrate. The extent of this area of ' unnavigable polar ice ' is charted in the map of the Arctic Regions (*The Oxford Atlas*, 1951, p. 9, Oxford University Press), and is reproduced in Fig. 6 (dark area). This area is surrounded by loose pack-ice which varies in thickness and extent with the seasons; during the winter pack-ice fills the Arctic Ocean; while in the summer it is loose with large areas of open water so that modern ice breakers are able to keep open a precarious passage for shipping along the north Asian coast and individual voyages have been made along the north American coast.

If the Arctic Ocean had been land the unnavigable area would certainly have become a permanent ice-sheet. For geographical reasons the unnavigable area is not concentric with the North Pole; on the average its present margin is 12° of latitude from the pole, which means an average latitude of 78°N. Taking into account the area filled with pack-ice covering the whole of the Arctic Ocean for the greater part of the year we shall not make any great mistake if we assume that the present ice in north polar regions is equivalent to an ice-cap with its margin along the 75°N parallel. It may be remarked that all the islands in this area are glaciated to sea-level, so would be part of the permanent ice-cap if the Arctic Ocean had been land.

Fig. 6 shows also the area in the northern hemisphere which we know from glacial remains was covered by an ice-cap at one time or another during the Pleistocene (light area). From geological evidence alone, we cannot say whether the whole of this area was glaciated at the same time; but there is evidence which will be discussed later that all the Pleistocene ice-caps reached to practically the same limits at their maxima; this was certainly the case of the Mindel, Riss and Würm glacial phases; so we may take the area shown in Fig. 6 as representing the north polar ice-cap at its maximum during at least three glacial phases. Owing to geographical factors this ice-cap also was not symmetrical about the North Pole. The average distance of the margin from the pole was 37° of latitude; so that its average latitude was 53°N. As this can only be an approximate figure, the more convenient figure 55°N will be adopted in what follows as the average latitude of the margin of the north polar ice-cap at its maximum during a glacial phase.

We are now in a position to discuss the changes in the extent of the north polar ice-cap with changes in the solar constant and to enter them in column (d) of Fig. 1. At the bottom of that column a scale of north latitude is added as abscissa and as ordinates we use again the scale of radiation in column (a). With these coordinates a curve has been constructed connecting the latitude of the margin of the north polar ice-cap, supposed symmetrical around the North Pole, with values of the solar constant in the following way.

It has been agreed above that the average latitude of the margin of the present north polar ice-cap may be taken to be 75°N and this is marked at B in Fig. 1, column (d) with coordinates : lat. 75°N, solar constant S_B. According to the solar radiation theory this ice-cap will grow as the solar radiation increases owing to increased precipitation until it reaches a maximum extent and will then decrease. If the radiation increases sufficiently the ice-cap will disappear entirely; but at what amount of radiation this would occur we cannot say. There is every reason to believe, however, that in each of the interglacial phases the ice-cap was as small as, if not smaller than, it is at present. Therefore there must be some point on the radiation curve in the upper part of our diagram at which the ice-cap returned from its maximum at lat. 55°N to its present size at lat. 75°N even if it did not disappear entirely. We arbitrarily mark this position at D with coordinates : lat. 75°N, solar constant S_D. We do not know the value of S_D except that it is the solar constant at which the ice-cap returns to its present size owing to increased solar radiation.

Somewhere between B and D the ice-sheet reached its maximum extent and this, as we know, was at lat. 55°N; thus the curve representing the extent of the ice-cap as it increases must reach the ordinate of lat. 55°N. As we have no evidence as to the radiation at which this occurs, except that it must be greater than S_B, and less than S_D, a simple free-hand curve has been drawn from B to D touching lat. 55°N at the point C.

Above D the ice-cap rapidly diminishes, as the radiation and consequent melting continue to increase, and finally disappears entirely; the curve is therefore produced beyond D to reach lat. 90°N in E, which signifies the total disappearance of permanent ice at sea-level at the North Pole. On the other hand below B the ice-cap continues to contract because with decreasing temperature the precipitation also decreases but it never ceases entirely; finally it will be little more than hoar frost in winter which will disappear in summer. At this point, which we mark A, there is no permanent ice-cap at the North Pole.

We have now drawn a curve which connects the extent of the north polar ice-cap directly with the solar radiation, column (a), and indirectly with the temperature and precipitation, column (b) and (c). This will be a convenient place to discuss an interesting point regarding the maximum extent reached by the ice-sheet in successive glaciations.

Wherever it is possible to determine the position reached by the ice-caps in Europe, both polar and Alpine, it is remarkable how near to one another were the margins of the last three advances at their maxima. The same has been found for the borders of the four principal drift sheets in central and eastern United States where the borders were not just near one another but were practically identical. Professor F. R. Flint (1953) who established this remarkable fact wrote in 1952 '*the broad parallelism and near congruence of the margins of the glacial drift sheets in central North America suggest that the climate controls of the origin and growth of the first ice-sheet were repeated during subsequent glaciations*' (p. 176) and remarks '*The cause, however, can be considered more effectually by the climatologist than the geologist*' (p. 167). I trust that Professor Flint will find the following explanation a satisfactory reply to his challenge.

We have already seen that when with increasing radiation the ice-cap has reached its maximum, at C in Fig. 1, there is a balance between the melting at the margin and the supply of ice from the ice-cap (see § 2). If now the radiation continues to increase both the temperature and precipitation will increase, but a decreasing proportion of the latter will fall as snow and the former will cause increased melting, the result being a contraction of the ice-cap along the curve C, D, E. On the other hand if the radiation decreases, the precipitation will decrease and in spite of lower temperature the ice-cap will contract and retrace its path along curve C, B, A and disappear at A. Thus a polar ice-cap does not continue to expand so long as the radiation increases and then to contract when the radiation decreases, or *vice versa*; it reaches its maximum extent when the solar radiation has the definite value Sc and then contracts whether the radiation continues to increase or commences to decrease. This value depends on two parameters (a) the shape of the Earth and (b) the chemical constitutions of the atmosphere.

As we shall see in § 8, according to the solar radiation theory the climate during the Pleistocene Period was governed by three oscillations of solar radiation from high values well above, to low values well below Sc, thus going through Sc twice in each oscillation. Therefore twice in each oscillation a polar ice-cap was formed which expanded to a maximum when Sc was reached and then contracted as Sc was passed, once when increasing and once when decreasing. Therefore in each glacial phase of the oscillation the ice-cap reached its maximum extent and each maximum had the same margin. This is the 'climate control of the origin and growth of the first ice-sheet' asked for by Professor Flint and it has been 'repeated during subsequent glaciations.' This however does not complete all the interesting results of the solar radiation theory.

As there have been no changes in the shape of the Earth or in the constitution of the atmosphere during the Pleistocene and probably none during geological time, the polar ice-caps can never have extended beyond lat. 55°N (approx) and will have extended to this latitude only when the solar constant has had the value Sc. It is also interesting to notice that there can be no permanent ice-cap when the solar constant is above Se or below Sa. Thus the values of the solar constant Sd, Sc and Sa have a peculiar interest as they fix the values of the solar radiation between which polar ice-caps can occur, Sa to Se, and the value at which the extent of the glaciation is at its maximum, Sc; they remain unchanged so long as the earth retains its present shape and constitution, they may therefore well be called ' terrestrial constants.' Unfortunately we do not know their numerical values, but I do not believe that it is beyond the wit of man to determine them. A corollary of this is that if evidence of extensive ice action at sea-level is found at any locality now between lat. 55°N and 55°S that land surface must have changed its latitude since it was glaciated.

Returning now to the construction of Fig. 1. It will be remembered that we have completed column (d) by constructing a curve relating the extent of the polar ice-cap to the solar constant; in column (e) we have now to represent how the climate during the glacial and interglacial phases responds to the changes in the solar radiation. It is usual at present to classify the climate in high latitudes, where alone glaciation at sea-level occurs, as either glacial or interglacial; but no definitions have been given. Generally speaking, periods during which there was extensive glaciation in regions where it does not occur today are said to have been in a glacial phase, and the regions affected by the ice are said to have had a glacial climate. The periods between two such glacial phases are said to be interglacial phases and the climate to be interglacial. It will be noticed that the conditions at the present time are used as the basis, and it is a suitable basis for we have quite recently (in geological time) passed out of a glacial phase; in fact at the present time there is quite an appreciable ice-cap at both poles, and there are ice-topped mountains in all latitudes, including the equatorial zone. While no one considers that we are living in an ice-age, with these relics of the last ice-age about us, and the knowledge that it is only relatively recently that the ice-cap withdrew from Scandinavia, it is clear that we are not far removed in time from the Würm glacial phase. We will therefore adopt the present time as being on the dividing line between a glacial and an interglacial phase. Looking at column (d) in which B represents the present time with an ice-cap extending to 75°N, we see that with increasing radiation the ice-cap increases, indicating greater glaciation than at present, hence a glacial phase is entered; as the radiation increases the ice-cap expands with increased glaciation, until the maximum of the ice-cap is reached at C; the ice-cap then extends to an average latitude of 55°N, advanced to 40°N in North America and retarded at lat. 60°N in North Asia, Fig. 6. From this maximum the ice-cap retreats until the average latitude of its margin is 75°N with the rest of the northern hemisphere free from permanent ice at sea-level. These are conditions similar to those of today, but with appreciably higher temperatures and much increased cloud and precipitation – the glacial phase is at an end. Thus when the extent of the ice-cap is between B and D, i.e. the radiation is between Sb and Sd the climate is glacial. This has been shown in column (e) of Fig. 1. We have now two quite definite limits for the glacial phase at B and D; this gives us a practical definition for a glacial phase : a glacial phase exists whenever the margin of the north polar ice-cap extends on the average south of latitude 75°N.

This is a satisfactory definition for the glacial phase; but it leaves us with two kinds of interglacial phases, (a) one with the solar constant greater than Sd, involving high temperature with much cloud and precipitation and (b) one with the solar constant less than Sb, involving low temperatures with little cloud and little precipitation. These two

interglacials will therefore be differentiated by describing them as the warm-wet interglacial and the cold-dry interglacial. Thus we have three kinds of world climate differentiated with respect to glaciation as shown in Table 1. These climates have been entered in column (e) of Fig. 1.

TABLE 1.

	Radiation	Ice-cap (margin)	Climate
(a)	greater than S_D	none or north of 75°N	warm-wet interglacial
(b)	between S_D and S_B	between 75°N and 50°N	glacial
(c)	less than S_B	north of 75°N or none	cold-dry interglacial

6. Pluvial and interpluvial phases

Pluvial and interpluvial phases are generally considered to be phenomena of low, if not equatorial latitudes; but if what has been said above under 'precipitation' (§ 4) is correct and precipitation in all parts of the earth increases and decreases with the solar radiation, all latitudes must be affected in the same direction if not to the same extent. In high latitudes, however, the effect of the ice-phase, which produces glacial conditions complicates the evidence of other climatic changes; for the effects of ice are so much more evident and lasting than those of water. On the other hand in low latitudes, which were not glaciated during the Pleistocene except on a few high mountains, there are only the effects of water to give a clue to past climates. Now the effects of changes in precipitation are much more marked in dry regions than in regions of high rainfall with their rain forests and luxuriant tropical vegetation; in consequence most of the information about past climates in low latitudes is derived from deserts and dry regions north and south of the equatorial rain band. In these dry regions the effects of excessive rain are much more evident than those of any decrease in the small normal rainfall; it is natural therefore, as prehistorians have always done, to describe past climates in low latitudes as pluvial or interpluvial. If what has been said above about glacial and interglacial phases is correct we have now to recognize that there are three glacial phases and only two pluvial phases. As this has not been recognized previously there has been a tendency for prehistorians to correlate all their pluvial periods with glacial phases and their interpluvial periods with interglacial phases which is impossible on the solar radiation theory. Our next object, therefore, must be to determine the relationship between the glacial and pluvial phases according to the solar radiation theory for entry in Fig. 1, column (f).

In 1930 Dr. Wayland defined a pluvial period in the following words : "*A period of geological signification during which rainfall was in general considerably heavier than in earlier times and than it is today over an area sufficiently large to be of some account in world events is called a pluvial period.*" Dr. Wayland tells me that this has not been entirely satisfactory, for there is a real need in African prehistory for a class of precipitation between the present rather dry conditions and the very wet conditions associated with the word 'pluvial.' To meet this need I propose to introduce four classes of precipitation as follows :

The present-day climate in low latitudes compared with previous climates, as judged by the extent of the water in lakes and water courses, is dry and if it became appreciably drier it would definitely be very dry; thus it is convenient to describe the present-day climate as being on the division between a very dry phase and a dry phase. We therefore produce the line through B, which represents present-day conditions, into column (f) to form the division between the very dry phase and the dry phase.

With increasing solar radiation the dry phase will change into a wet phase and it is convenient to make this change in nomenclature coincide with the maximum of the ice-cap; hence the line through C in column (d), which represents the maximum of the ice-cap, has been produced into column (f) to mark the division between the dry phase and the wet phase. With a still further increase in radiation and precipitation the wet phase change into a very wet phase. As the very dry phase occurs when the radiation is less than during the glacial phase it will be natural to make the very wet phase occur when the radiation is greater than in the glacial phase, that is above D in column (f). It will be noted that if we combine the wet and the very wet phase into one phase and call it the pluvial phase, then the dry and very dry phases can similarly be combined into an interpluvial phase. In this way we have four precipitation phases which are directly related to the three glacial phases and we meet the wish of Dr. Wayland for a climatic phase, wetter than today, but not so wet as a pluvial phase.

We can now collect the relationships between the climates and solar radiation in the following table.

TABLE 2. SOLAR RADIATION AND CLIMATES

Solar constant	Latitude of margin of ice-cap	Glacial phases		Pluvial phases
Greater than S_D	North of 75°N or no ice-cap	Warm-wet interglacial	Pluvial	Very wet
Between S_D and S_C	Between 75°N and 55°N	Glacial	Pluvial	Wet
Between S_C and S_B	Between 55°N and 75°N	Glacial	Interpluvial	Dry
Less than S_B	North of 75°N if any	Cold-dry interglacial	Interpluvial	Very dry

7. CHRONOLOGY

In our discussion so far we have only considered how the elements of the climate, temperature, glaciation and precipitation, are related to one another. We have now to study the actual climate as it oscillated between glacial and interglacial phases and between pluvial and interpluvial phases. The first need for such a study is to have a chronological record of the climate, which can only be obtained from geological evidence. My first concern on undertaking this further study of the Pleistocene climate was to obtain an authoritative statement of geological opinion on this matter.

I was delighted therefore to come across the chronological chart of the glacial and pluvial phases of the Pleistocene which forms an appendix to '*Man the Tool-maker*' a handbook published by the British Museum (Natural History) and written by Dr. K. P. Oakley who himself has worked on glacial deposits in Europe and on pluvial deposits in East Africa. I was still further pleased when on meeting Dr. Oakley he told me that he was preparing a new (the third) edition of '*Man the toolmaker*' for which he had revised the chronological chart to bring it up-to-date, an advance copy of which he kindly gave to me. I have therefore decided to base my work, so far as chronology is concerned, on this chart, fully realising the limitations of all such estimates. The pluvial glacial correlations in Oakley's chart were mainly based on the use of tool-traditions as time-lines.

The relevant parts of Oakley's chart are reproduced in the upper half (a) of Fig. 2. The first three lines are copied from Oakley's chart. In the first line the Pleistocene Period is divided into Lower, Middle and Upper divisions; these divisions however are not drawn to scale, " the earlier phases having been telescoped for convenience " to quote Dr. Oakley. In the second line the pluvial phases are indicated by horizontal lines ending in short vertical strokes, and in line 3 the glacial phases are indicated in a similar manner. Oakley does not attach names to these phases in his chart, but says in his introduction : " The glacial periods shown on the chart are generally regarded as corresponding with the advances of ice in the Alps which have been named Günz, Mindel, Riss and Würm . . . . . The pluvial periods recognized in Africa are now named Kageran, Kamasian, Kanjeran (Upper Kamasian) and Gamblian." I also do not think it is advisable to label these phases in the diagram, at least not until we have discussed them further, and so in Fig. 2 I have marked the pluvial phases in line 2 : P1, P2, etc. and the glacial phases in line 3 : G1, G2, etc. for convenience of reference.

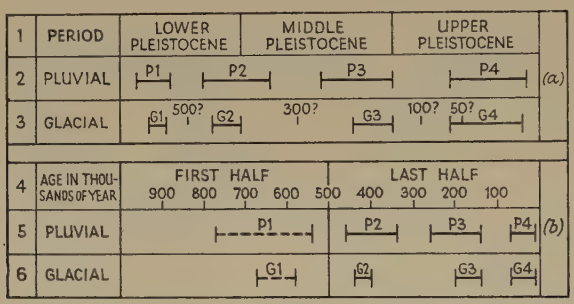

Figure 2. Pluvial and glacial phase (Oakley) : (a) telescoped scale, (b) true scale.

As already stated the time scale in the original diagram, line 1, is not linear, the earlier phases having been telescoped. For our purpose a true time scale is essential, the rate at which the climate changes being an important element in the solution of our problem. Oakley takes one million years as the time occupied by the whole of the Pleistocene Period. He also enters at the middle of each interglacial an estimate of the time in thousands of years which has elapsed since that epoch; to stress that these are only estimates he adds a query mark against each. These estimated times, with their queries, have been reproduced in line 3 of Fig. 2. Although these are only estimates the position they occupy in the diagram do give the scale of time Oakley had in mind when preparing his chart. It was a simple matter to plot these estimates against their positions on the chart to recover the true time scale. This proved to be nearly linear from 0 to 500,000 years ago (last half of the whole period) and then the time scale decreased at a rapidly increasing rate so that the whole of the first half of the Pleistocene, 1,000,000 to 500,000 years ago, is telescoped into a length only one-sixth of that of the whole period.

In the lower half (b) of Fig. 2 lines 4-6, the positions of the pluvial and glacial phases shown in the upper half have been replotted against the linear time scale thus recovered, the whole period of one million years being divided equally into units of one thousand years.

Having decided to use Oakley's chart as the chronological foundation of our discussion does not release us from the duty of examining it carefully and pointing out any limitations. The first thing I noticed on transforming the time scale was that with the original scale, Fig. 2, lines 2 and 3, the phases appear to be more or less evenly distributed over the whole of the Pleistocene Period; while with the linear time scale, lines 5 and 6, there is only one of each phase in the first half of the Period, the other three of each being crowded into the second half.

Now this might be due to a real increase in the frequency of the glacial and pluvial phases in the second half of the Period; or to the geological remains from the first half of the Period having become indistinct owing to subsequent weathering and largely removed by subsequent advances of the ice. I am inclined to the second explanation by noticing that P1 in line 2 and G1 in line 3 are entered centrally in the very small space given in Oakley's original chart to the first half of the Period. With the very small and rapidly changing time scale used in this part of the original chart it would be quite impossible to fix any period with even approximate accuracy. In the future discussion, therefore, I shall not use P1 and G1 except as indications that there is evidence of both pluvial and glacial phases sometime during the first half of the Pleistocene; they have therefore been indicated in lines 5 and 6 by broken lines.

8. The climate curve

If we plot a curve on a diagram with rectangular coordinates, taking radiation (solar constant) for ordinates and time for abscissa, then each point on the curve represents a definite solar constant which specifies a pluvial and a glacial phase according to Table 2. Inversely if we are given the times of the beginnings and endings of a sequence of glacial or pluvial phases, then it is only necessary to plot the solar constant at which these phases begin and end from Table 2 against the relevant times and join the points by a suitable line to have a curve from which any of the factors represented on Fig. 1 can be read. Such a curve can be called a climate curve and we will now proceed to construct one for the Pleistocene Period from the data contained in Figs. 1 and 2. In preparing the curve we shall use only the glacial phases and if this is successful the pluvial phases indicated by the curve will be compared with those observed; if they agree our problem will be solved, if not, either the observations or the theory must be at fault.

On Fig. 3 the following data are given :

Lines 1 and 7 : a scale of time covering the million years of the Pleistocene Period;
line 2 : the three divisions of the Pleistocene Period as used by Oakley*;

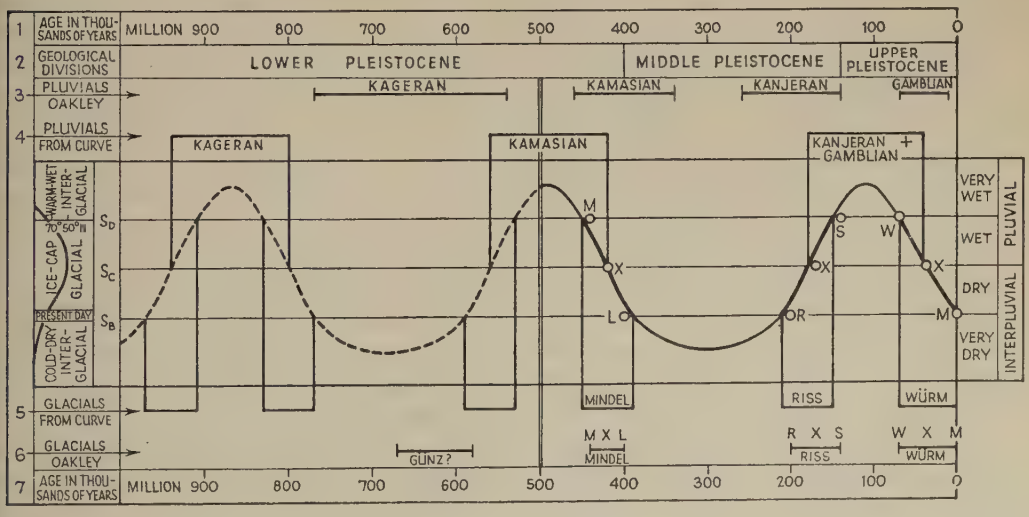

Figure 3. The climate curve.

* The base and top of the Middle Pleistocene are here drawn in accordance with the practice adopted in a book by F. E. Zeuner, *The Pleistocene Period* (Ray Society, London, 1945), p. 175; but some authorities on the Pleistocene now shift the base down to include the Mindel and the top down to exclude the Riss (K.P.O.).

line 3 : Oakley's pluvial phases from Fig. 2 line 5 plotted against the time scale in line 1;
lines 4 and 5 : these will be described later;
line 6 : Oakley's glacial phases from Fig. 2 line 6.

The diagram :

The abscissae are times on the scale in lines 1 and 7.
The ordinates are radiation (solar constant) on scale of S in Fig. 1 (a).
The limits of the glacial phases from Fig. 1 (b) are shown against the solar constant on the left of the diagram.
The limits of the pluvial phases from Fig. 1 (f) are shown on the right of the diagram.

(*a*) *The construction of the climate curve*

The curve is to be constructed using only the data we have about the glacial phases. The periods of Oakley's glacial phases from Fig. 2 (b) are entered in line 6. We have agreed not to use the first glacial phase, G1, which Oakley shows in the Lower Pleistocene because its position and extent are too indefinite. This leaves us with three glacial phases, all in the second half of the Pleistocene Period. These are so recent that there can be no doubt that they are the Mindel, Riss and Würm glacials; we will, therefore, enter these names in line 6 of the diagram. For convenience of description the begining and the end of each phase is marked with the initial and final letters of its name, and X is entered in the middle to represent the time of the maximum extent of the ice-cap during the phase.

We now draw lines parallel to the base line across the whole extent of the diagram at S_D and S_B : which are the solar constants dividing the glacial from the interglacial phases according to Fig. 1 (e). All points above the S_D line are in the warm-wet interglacial phase, all points between the S_D and S_B lines are in the glacial phase and all points below the S_B line are in the cold-dry interglacial phase. The glacial phase of the climate changes whenever the solar constant crosses one of these two lines. A third line is drawn across the diagram at Sc. It is at this solar constant that, according to Fig. 1 (d), the ice-cap reaches its maximum extent during each glacial phase.

To plot the curve the absolute coordinates of at least one point on it must be known. There is only one point fulfilling this requirement, namely the present epoch for which the coordinates are : solar constant $S_B = 1{\cdot}95$ cal cm^{-2} min^{-1}; time = 0 on the time scale in line 1; this point is also by definition the end of the Würm glaciation; its position on the diagram is marked M because it corresponds with M, the end of the Würm glaciation in line 6.

The positions corresponding with the maxima of the ice-caps are now marked X. These are all on the Sc line where it meets the ordinates erected at the three Xs in line 6. The begining and ending of each glacial phase has now to be marked on the diagram. We start at the point corresponding with the present epoch which we have already plotted as M. The next point, the maximum of the ice-cap in the middle of the Würm glaciation, has already been plotted at X. To get out of the glacial phase from this point we must cross the ordinate at W (the begining of the Würm glacial phase) where it meets either the S_D line or the S_B. Of these S_D is the one we will take for a number of reasons which it is unnecessary to go into here. This point has been marked W on the diagram. Passing on to the Riss glaciation, the glacial phase is entered at S; the maximum of the ice-cap is reached at X and the interglacial phase entered at R. Similarly passing on to the Mindel glaciation the glacial phase is entered at L and left at M.

We have now marked nine points on the diagram each representing the climate at the corresponding epoch. When these points are joined by a succession of straight lines the general shape of the climate curve is revealed as a wave curve with two maxima and one minimum in the period covered. Trial and error then gives the simple curve shown in the diagram which is completely symmetrical and passes through or very near each of the nine plotted points.

This, however, is not the only smooth curve which can be drawn through these nine points; but as the climate curve can only enter a glacial phase at either of two points (where the ordinate at the beginning of a phase crosses the SD and the SB lines) and leave at two other corresponding points at the end of the phase, the number of curves is limited. I have tried them all; but the one shown is the most satisfactory; for it makes fewer changes in direction and fits the pluvial phases better than any other.

The curve represents the glacial evidence during the second half of the Period so well that it makes one regret that the data for its extension into the first do not exist. It is difficult to believe, however, that there is no physical cause behind the climate curve. If there is, it would be even more difficult to believe that such a cause could have developed and become so pronounced and regular for half a million years without having taken some considerable time to build up. I have therefore extrapolated the curve into the first half of the Pleistocene, but used a broken line to signify that it is not based on observations and therefore should not be given the same weight as the curve in the recent half of the Pleistocene.

We have now drawn a climate curve for the whole of the Pleistocene Period based on the glacial phases only; it remains to compare it with the glacial phases observed by geologists and summarized in Oakley's chart.

(*b*) *The glacial phases*

The sections of the curve between the two horizontal lines at SD and SB represent the glacial phases and have been thickened. Vertical lines have been drawn downwards from the beginning and ending of each of these glacial sections of the curve and extended for a short distance below the base line where they are joined together by short horizontal lines, Fig. 3 line 5. In the whole period there are six of these short horizontal lines, each one representing a glacial phase of the climate. When the three in the last half of the Period, representing the Mindel, Riss and Würm glaciations on the climate curve, are compared with the corresponding short lines in line 6, representing Oakley's glacial phases it will be seen that the correspondence is good.

The position is not so simple in the first half of the Pleistocene Period. Penck and Brückner recognized glacial action in the Alps before the Mindel and gave the name Günz to it (Penck and Brückner 1909). Naturally whenever and wherever the Mindel glaciation has been recognized and evidence of an earlier glaciation has subsequently been found the latter has been assumed to be Günz. The climate curve, however, shows three glaciations before the Mindel and although at first I was disposed to consider the third of these to be Günz I now think that it would be wiser at this stage not to assign names to these three glaciations, but to consider Günz, like Oakley's G1, to indicate evidence of glacial action some time during the Lower Pleistocene (see p. 471). The name Günz can then be assigned to any one of the three when future investigations justify it. The climate curve now represents the known glacial phases in the last half of the Pleistocene Period and indicates when earlier phases may be expected in the first half.

9. The climate curve and the pluvial phases

In the previous section a curve, which I have called a climate curve, has been constructed for the whole of the Pleistocene Period and in preparing it only observations of the glacial and interglacial phases have been used. The pluvial and interpluvial phases are also governed by the solar radiation as completely as the glacial phases and the relationship is shown in Fig. 1. We are therefore in a position to add the pluvial phases to the climate curve in Fig. 3 by use of the scale in column (f) of Fig. 1 which has been reproduced on the right of the diagram in Fig. 3.

Whenever the climate curve in Fig. 3 crosses the Sc line the climate changes from the pluvial phase to the interpluvial phase or vice-versa. This happens six times on the diagram and from each crossing point vertical lines have been erected and produced for a short distance beyond the top line of the diagram. The vertical lines between which the curve is above Sc have been connected by short lines. These short lines mark the pluvial phases according to our definition. We have now, from a climate curve constructed by using glacial phases only, derived the sequence of pluvial phases shown in line 4 and our next problem is to correlate them with the observed pluvial phases represented in line 3.

The four observed pluvial phases in line 3 are assigned by Oakley to the Kageran, Kamasian, Kanjeran and Gamblian pluvials respectively. Assuming for the purposes of the discussion that the pluvial sections of the curve, indicated in line 4 of the diagram, are correct and correspond with actual pluvials, then there can be little doubt that the first pluvial on the curve is the Kageran, for this is the only pluvial recognized by geologists in the Lower Pleistocene on geological and cultural evidence; '*In the valley of the Kafu river, Wayland first found pebble-tools of what is probably the earliest known culture in the world, the Kafuan pebble culture. These tools were associated with Kageran deposits of Lower Pleistocene age*' : (Cole p. 44); '*the Kageran pluvial at the very begining of the Pleistocene*' : (Leakey 1953, p. 66).

The second pluvial section of the curve is even more clearly correlated with the Kamasian Pluvial. In the first place the pluvial on the curve is very near in time to Oakley's second pluvial, the Kamasian pluvial. In the second place the Kamasian pluvial was followed immediately by a very arid period : '*The upper beds at Rawi were formed during this arid period; they contain abundant remains of fish which evidently died as Lake Victoria diminished. The only fauna which survived was mud-fish and lung-fish*' (Cole p. 47). This arid period following the Kamasian pluvial is clearly represented in the diagram by the dry and very dry sections of the curve immediately following the second pluvial. In the third place there are cultural links between the Kamasian pluvial and the Mindel glacial phase : '*Primitive pebble-tools were replaced by the . . . . . hand-axe culture in the very beginning of the Mindel glacial phase in Europe and in early Kamasian times in Africa*' (Cole p. 123). It will be seen that the pluvial section of the curve we are discussing overlaps the Mindel glacial phase. There can be little doubt that the second pluvial section of the curve corresponds to the Kamasian pluvial.

We now come to a difficulty. According to line 3 of the diagram there are two more observed pluvials, the Kanjeran and the Gamblian, and only one more pluvial indicated on the curve. If the curve is really correct and the first two pluvial sections represent the Kageran and Kamasian there appears to be no alternative to assigning the third pluvial section of the curve to the Kanjeran. We should then have to add another wave to the curve to provide for the Gamblian. That is obviously impossible for the curve has already reached the present day; nor can it be added at the beginning of the curve for that is fixed by the physical correspondence of the two first pluvials on the curve with the Kageran and Kamasian pluvials.

There would be no solution to this problem if the Kanjeran and the Gamblian were pluvial phases similar to the Kageran and Kamasian with an interpluvial between them similar to the one following the Kamasian pluvial with its diminishing Lake Victoria and its dying fish; but there is nothing like that recorded. On the other hand it is only necessary to assume that the Kanjeran and Gamblian pluvials are really one – with possibly a short, temporary decrease in the precipitation near the middle – and the whole difficulty disappears. I do not intend to argue this point; it is much too vital to be solved by any other means than observations in the field. Judging by the splendid work which has been done in prehistory in East Africa since Wayland discovered, in 1919, the first stone implements recognized as such in Uganda it will not be long before the question is settled. In the meantime I content myself by noting that in Mrs. Cole's book the evidence for an appreciable break between the Kanjeran and Gamblian pluvials is very scanty and far from convincing. I shall therefore proceed with the discussion on the assumption that the Kanjeran and Gamblian together form the last pluvial .

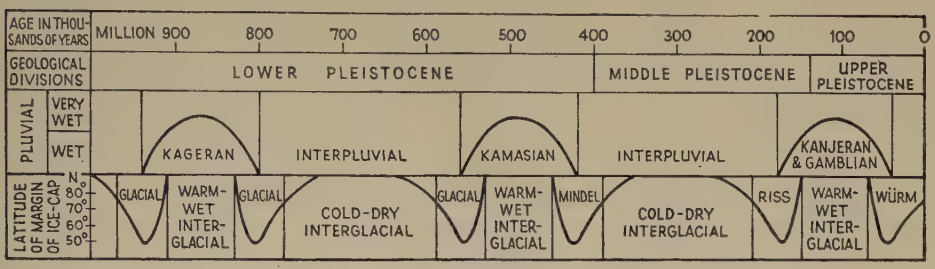

Figure 4. Pluvial and glacial phases (from climate curve).

10. The relationship between the pluvial and glacial phases

The relationship between the pluvial and glacial phases is not easily recognized from Fig. 3. Another diagram, Fig. 4, has therefore been prepared to bring out this relationship more clearly. It will be seen from Fig. 3 that each pluvial phase on the climate curve extends from the maximum of each ice-cap which is associated with increasing radiation to the maximum of the next ice-cap associated with the subsequent decrease of the radiation. During this interval of time the radiation and with it the precipitation rises to a maximum and then decreases. These periods of maximum radiation and precipitation have been copied into the upper half of the new diagram, Fig. 4, to represent the precipitation during the pluvial phases. Between the pluvial phases the climate is interpluvial and for clarity the intensity of the precipitation during these periods has been omitted; all that is necessary to remember is that in all the interpluvials both the precipitation and the mean annual temperature are less than during the pluvial phases.

The glacial phases are shown in the lower half of Fig. 4; these have been plotted in the following way. On the left-hand side of the diagram in Fig. 3 the curve relating the latitude of the ice-cap to the solar constant in Fig. 1 has been reproduced. Each point on the climate curve corresponds with a point on the latitude curve and these points have been read off and plotted in the lower half of Fig. 4, the resulting curves representing the advances and retreats of the ice-cap during the glacial phase. From these curves we see that each glacial phase consists of an advance and retreat of the polar ice-cap; the glacial phases occur in pairs, the pairs being separated by long interglacials; while the glacial phases forming each pair are separated by short interglacials. The climate is not the same in the long as in the short interglacials, the long ones being cold-dry and the short ones warm-wet.

Looking at Fig. 4 as a whole we now see that instead of the simple direct correlation of each pluvial phase with a glacial phase and each interpluvial phase with an interglacial phase as generally supposed by prehistorians the relationship is much more complicated. According to the diagram while each pluvial phase begins and ends during a glacial phase its centre does not coincide with a glacial phase; on the other hand the centre of each interpluvial does coincide with an interglacial. It is this complication which has made it practically impossible to correlate the pluvial with the glacial phases from field observations alone.

11. Conclusion

We have now constructed from meteorological and geological data a simple curve to represent the changes in world climate during the last million years (Fig. 3).

On comparing this curve with the glacial phases discovered by geologists we find that of the four glacial phases known to have occurred in North America and Europe, three, the Mindel, Riss and Würm, are represented on the curve in the second half of the Pleistocene Period; while the curve indicates three more glacial phases in the first half, one of which is without doubt the Günz.

Of the four pluvial phases known from East Africa the first two, the Kageran and Kamasian, are correlated with the first two pluvials indicated on the climate curve during the Lower and Middle Pleistocene. The curve also indicates one more pluvial phase in the Upper Pleistocene; the geologists, however, find evidence of two pluvials, the Kanjeran and Gamblian, in this period of time. Reasons are given for suggesting that the two latter pluvials are really one, the break between them being only a minor feature not comparable with the other two earlier interpluvial phases. If further field observations confirm this suggestion the climate curve will represent all the known pluvials as well as all the known glacial phases. The solar radiation theory will then not only give us an explanation of the cause of the Pleistocene climates, but also considerable knowledge of the changes in solar radiation during the last million years.

Acknowledgments

I wish to express my thanks to Mrs. Sonia Cole and Dr. K. P. Oakley of the British Museum, London; to Dr. L. S. B. Leakey, Director of the Coryndon Museum, Nairobi; to Sir Charles Normand, Oxford; and to Dr. E. T. Eady of the Imperial College of Science, London, who read and commented on an early draft of this paper. Their criticisms and suggestions have contributed largely to the paper in its present form and I doubt whether it would have been completed without their encouragement. I am deeply grateful to them all.

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Appendix

In the foregoing paper the solar-radiation theory has been discussed in a purely qualitative manner, the only numerical values being the present day values of the solar constant, the mean temperature of the earth and the total amount of precipitation; all other quantities have had to be described in words or represented by symbols. The whole idea that an ice-age with great polar ice-caps and heavily glaciated mountains was the result of an appreciable increase of the temperature of the Earth as a whole due to a warmer sun is so foreign to our way of thinking (in spite of a refrigerator in nearly every home), that it is often instinctively rejected as being against ' common sense.'

Two well-known meteorologists to whom I sent an early draft of the paper asked the same question : " Could the present ice-cap extend to lat. 55°N, the zonal temperature of which is at present above the freezing point, if the mean temperature of the earth increased " ?

A theory unchecked by numerical values is in any case very weak and when it runs so contrary to our preconceived ideas some numerical evidence that the scheme is at least not inherently impossible becomes imperative. I therefore continued my further studies and obtained some numerical values of temperature at maximum glaciation which I add to my main paper as an appendix to indicate that they are not an essential part of the solar radiation theory, but only a support.

For this study a simplified model of the Earth was adopted in which land and sea is evenly distributed, with all the land at sea-level, so that the ice-cap is always concentric with the pole and its margin is always along a circle of latitude. With this simplification both temperature and ice effects are symmetrical about the pole.

The present distribution of the sea-level temperature of the Earth, as represented by the zonal temperatures for January and July, was taken as the starting point and then an increase in the solar radiation which resulted in a glacial phase with a polar ice-cap at its maximum extent was assumed. This involved, on the radiation theory, an increase in the temperature at the equator of an unknown amount. An increase of 2°C was assumed and then on general physical and meteorological principles the probable changes in the zonal temperatures of all latitudes from the equator to the north pole were deduced. Then taking other temperature increases at intervals of two degrees the process was repeated until the general effect of increasing the radiation to the maximum of an ice-cap and beyond could be seen.

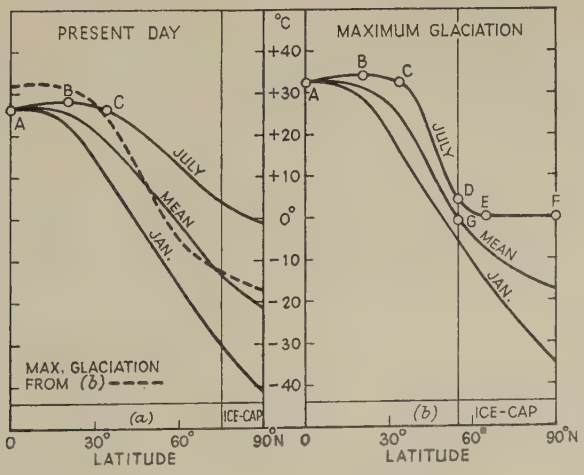

Figure 5. Zonal temperatures.

The results of the study are exhibited in Fig. 5 (a) and (b). In Fig. 5 (a) the present day mean temperatures along circles of latitude (zonal temperatures) as calculated by Meinardus (Hann – Suring 1926, p. 848) are given for the northern hemisphere for the months of January, July, and a third curve is added for ½ (Jan.-July) to represent the year as a whole. These three curves represent the temperature distribution with the present solar constant. From the preliminary experimental curves it had been found that at the maximum of a glacial phase an increase of 6°C in the temperature at the equator in all seasons best fitted the facts. This increase was adopted in the further discussion.

Using the present day distribution of the zonal temperatures, Fig. 5 (a), as a basis and applying an increase of six degrees at the equator at all seasons three corresponding curves, Fig. 5 (b), were derived for the zonal temperatures at maximum glaciation in the following manner :

January temperatures. At the maximum of a glacial phase the ice-cap extends to the position shown in the map, Fig. 6, which is represented on our model by a polar ice-cap with its margin along lat. 55°N.

At the present time during January every year the whole of the northern hemisphere from the pole to lat. 50° (approx), with the exception of the North Atlantic, is covered with a layer of snow at a temperature well below the freezing point, and the rest of the hemisphere is free from snow. These are practically the same conditions as during January at the maximum of a glacial phase (Simpson 1934, Figs. 5 and 6). We should therefore expect that the only effect of an increase of radiation during January would be to raise the temperature in all latitudes, somewhat more in equatorial than in Polar latitudes. Neglecting this latter difference we may take as a first approximation, that in all latitudes the January zonal temperatures were increased by the same amount. The curve for January at the maximum of a glacial phase, Fig. 5 (b) has therefore been drawn six degrees higher than the January curve in Fig. 5 (a).

Result : The January zonal temperature was 6°C higher than at present in all latitudes.

July temperatures. In July at the height of a glacial phase the conditions of the area represented on our model as extending from the pole to lat. 55°N were radically different from what they are at present. For then all this great area was covered by a thick sheet of snow and ice, while at present the ice-cap on our model extends only from the pole to lat. 75°N, and the rest is free from permanent ice. The thermal properties of an ice surface are very different from those of a land surface free from snow, the chief amongst which, for our purpose, is that the temperature of the snow surface can never rise above the freezing point. If we look at the temperature over the present ice-cap in Fig. 5 (a) we see that at the north pole the July temperature is one degree below freezing point; but this is not an observed temperature; it is however sufficiently near the freezing point to support the conclusion that the summer temperature over an extensive ice-sheet cannot be appreciably above the freezing point.

There would be no polar ice-cap if the annual snowfall on its surface did not exceed the annual loss through evaporation and melting. This excess causes the cap to thicken and to spread outwards under gravity; in doing so it reaches latitudes with higher temperatures and consequent increased melting until finally a position is reached where the annual marginal melting is equal to the annual outward flow of the ice-cap. From Fig. 5 (a) we see that on our model the present ice-cap reaches lat. 75°N where the July temperature is 4°C. These figures can only be approximate for neither the limit nor the temperature of the present ice-sheet is definitely known, so in default of better information we will adopt them and apply them to the much more extensive ice-cap of the glacial maximum. In Fig. 5 (b) therefore the July temperature is shown to be 4°C at the margin of the ice-cap in lat. 55°N, falling to 0°C at lat. 66°N, and then remaining at the freezing point to the

pole. These three points are marked D, E and F respectively in Fig. 5 (b). Thus the curve D, E, F is part of the curve for the July temperature at the maximum of a glacial phase.

When we examine the present July temperatures in the equatorial zone, Fig. 5 (a), we notice that the zonal temperatures at the equator, (A) and at 35°N (C), are the same, 26°C, with a slight rise of 2°C between them at lat. 20°N, (B). This distribution is due to the north declination of the sun in July and will not change with a change in the solar constant. We have assumed that the equatorial temperatures at maximum glaciation was in all seasons 6°C higher than at present; we therefore add 6° to the present July temperatures and obtain (A) lat. 0°, 32°C; (B) lat. 20°N, 34°C and (C) lat. 35°N, 32°C and plot them in Fig. 5 (b) to give the equatorial section of the July curve. If now we connect C and D by a smooth freehand curve we obtain a complete curve for the July zonal temperature from the equator to the north pole at the maximum of a glacial phase.

Result. From the equator to lat. 35°N the July zonal temperatures were 6°C higher than at present, then decreased to no change at lat. 45°N.

From lat. 45°N to the north pole the July zonal temperatures were lower than at present, the difference being greatest, 12°, at the margin of the ice-cap in lat. 55°N and decreasing to no change at the north pole.

Mean annual temperature. The mean annual (zonal) temperatures for all latitudes have been calculated as the mean of the January and July values and plotted in (a) and (b) of Fig. 5. The mean curve for the maximum glaciation in Fig. 5 (b) has been replotted as a broken curve in Fig. 5 (a) for convenience of comparison.

Result. From the equator to lat. 45°N the mean annual zonal temperature was higher than at present, the increase being 6°C at the equator decreasing to no change at lat. 45°N.

From lat. 45°N to lat. 70°N the annual zonal temperature was lower than at present, the greatest decrease being 3°C at the margin of the ice-cap in lat. 55°N.

From lat. 70°N to the north pole the annual zonal temperature was higher than at present, the largest increase being 3° at the north pole.

Reviewing the above we find that the greatest change in the zonal temperature at the maximum of the glaciations compared with those of the present time was an increase of 6°C in all seasons at the equator and in all latitudes during January; and the greatest decrease was 12°C during July at the margin of the ice-cap in lat. 55°N.

It is unnecessary to consider in detail the further changes as the radiation continues to increase. It is sufficient to point out that at the maximum of a glaciation the mean annual temperature at the margin of the ice-cap, G in Fig. 5 (b) is already practically at the freezing point; as it rises further with increasing radiation the melting at the margin will continue to increase until it exceeds the outward flow of ice when the margin must retreat until the ice-cap has entirely disappeared.

According to the solar radiation theory the solar radiation varies in the cycles represented by the radiation curve in Fig. 3; but the temperature changes we have been considering are all confined to the sections of the radiation curve between the horizontal lines through S_B (the radiation at the present time) and S_C (the radiation at maximum glaciation). This section is approximately only a third of a complete swing from minimum to maximum of the radiation. It would however be wrong to assume that the total temperature variations would be three times as great as those we have discussed. The temperature changes are not linearly associated with the radiation changes; other factors : ice, clouds, wind, etc., have to be taken into account and these make the problem quite insoluble by the descriptive method used in this appendix. I shall not, therefore, take the discussion any further; but will be content if it has shown that the formation of an ice-cap, extending on the average from the north pole to lat. 55°N, is quite consistent with an appreciable rise

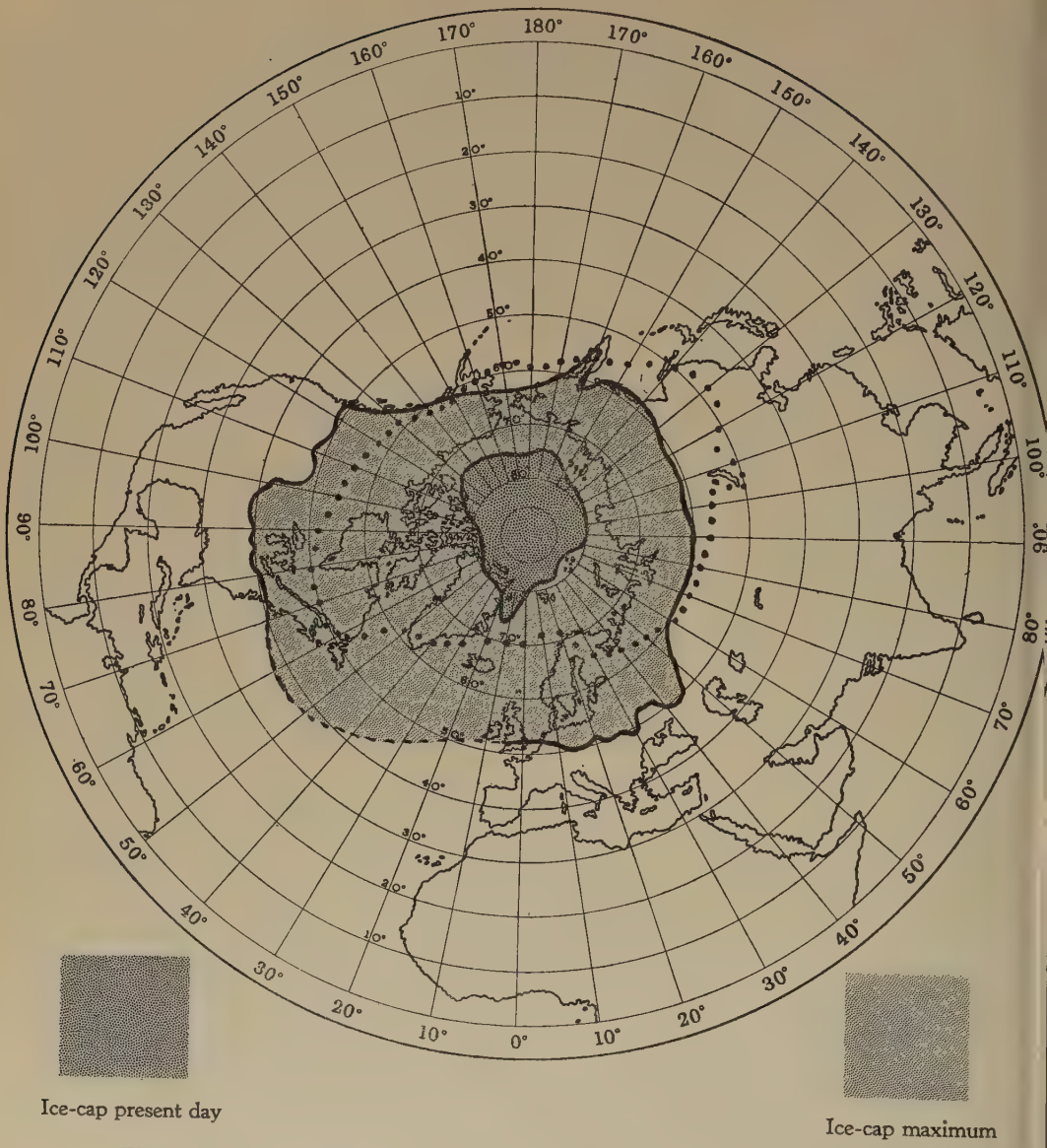

Figure 6. Polar ice-caps. . . . . mean annual isotherm 0°C, – – – – present day.

in the solar radiation. I must however stress again that the temperatures derived in this appendix are at best only approximations and may give little more than the orders of magnitude; but I do maintain that they show that the solar radiation theory is not inherently impossible.

Before closing I wish to call attention to an interesting outcome of the preparation of Fig. 6. On this map a dotted curve has been added showing the present position of the isotherm for 0°C mean annual temperature. Comparing this curve with the limits of the unnavigable area in the Arctic Ocean which we have adopted as representing the present polar ice-cap, it will be seen that they are separated by an extensive area. Now in this area the ground is necessarily frozen up to a few inches of the surface all the year round, and yet there is no ice-cap. From the beginning of my study of Pleistocene climates I

have considered this as good evidence that it is not the temperature which is the predominating factor in limiting the extent of the north polar ice-cap, but the want of sufficient snowfall in the winter to meet the summer evaporation and melting.

On the same map is shown the boundary of the ice-cap at the maximum of the glacial phase which we have represented on our model by an ice-cap with its margin along lat. 55°N. It will be seen from Fig. 5 (b) that the annual zonal temperature at this latitude was then 0°C. Thus at the maximum of the ice-cap, just when it is about to retreat because of increasing radiation, it completely fills the area in which the mean annual temperature is then below 0°C. Can this be explained otherwise than that at the maximum glaciation the snowfall was much greater than at present as required by the solar radiation theory ?

Discussion

Discussion by correspondence of Sir George Simpson's paper ' Further studies in world climate ' was invited. The contributions which follow were received in May and June 1957.

Dr. L. S. B. Leakey (Coryndon Museum, Nairobi) : The latest attempt at glacial-pluvial correlation which Sir George Simpson has set out in his paper on world climate is of particular interest to me because further work on pluvial problems has been undertaken and the evidence re-examined during the last few years, and the new ideas which are set out in the paper seem likely to fit in very well with the new interpretation of the evidence that is beginning to crystallize in East Africa.

Until recently, I have considered that there were four major pluvial periods during the Pleistocene of East Africa : the Kageran, the Kamasian, the Kanjeran and the Gamblian, and I have for some years now tentatively equated these with the Günz, the Mindel, the Riss and the Würm glaciations of the European scheme. The geological evidence for a really dry interpluvial both between Kageran and Kamasian, and between the Kamasian and Kanjeran is excellent, but the break between the Kanjeran and Gamblian has never been so clearly demonstrated on purely geological evidence and, indeed, was based upon rather different forms of evidence. It had, indeed, been assumed that the evidence for the dryness at this time had largely been obliterated by the tectonic activity which was so violent at this period.

It should, moreover, be noted that the Kamasian has a very definite break with two major peaks of increased precipitation and there is some evidence also that the Kageran had two peaks and one drier phase.

It now begins to seem possible that we in East Africa have only had three major pluvial periods, the Kageran, the Kamasian and a combined ' Kanjeran-Gamblian,' but that each of the three had an oscillation at about the mid-point which separated two major peaks. Moreover, there is, of course, ample evidence that during the Gamblian pluvial (which we now perhaps ought to call ' the Gamblian part ' of the combined Kanjeran-Gamblian pluvial) there were three further distinct smaller peaks of precipitation, each separated by a short phase of lesser rainfall.

If the hypothesis now set out by Sir George Simpson is accepted and if sufficient evidence can be obtained to confirm the presence of three glacial episodes prior to the Mindel, during the Pleistocene, then our East African evidence will fit in very well with the views set out in the paper under discussion.

Moreover, if, as a result of this modification, the Mindel glacial is only equated with a second maximum of the Kamasian pluvial, it would make the correlation of our African stone-age cultures with those of Europe much more readily intelligible. For instance, it would mean that the later stages of the pebble-tool culture, or Olduwan, which occurs in Bed I or Olduvai, would no longer have to be regarded as contemporary with the Mindel glaciation, but with the preceding glacial episode.

On the evidence at present available, the Kageran pluvial, which I think has two peaks, was probably of a less intensity than was the Kamasian or the combined Kanjeran-Gamblian. It should also be noted that the interpluvial between the Kamasian and the Kanjeran was probably much more intense than that between the Kageran and the Kamasian, or at least of longer duration.

Reverting, for the moment, to the question of pluvial-glacial correlation, it is, I think, increasingly obvious that what we used to call the Gamblian pluvial – with its three maxima – represents the Würm glacial. Therefore we start from that point in time and work backwards;

there is little alternative but to suggest that what we term the Kanjeran pluvial is the equivalent of the Riss glacial. If it is accepted that the evidence put forward by Sir George Simpson necessitates treating one pluvial period as being the equivalent of two glacials and one interglacial then we can only treat the dry period between the Kamasian and the Kanjeran as representing the major interglacial between the Mindel and the Riss, while the Mindel would become the equivalent of only the second maximum of the Kamasian pluvial.

Dr. C. C. Wallén (Institute of Meteorology, Stockholm) : I have been very much interested in reading the new article by Sir George Simpson on his glaciation theory. From what we know geologically about the ice-ages and the pluvial periods there seem not to be any strong arguments against the author's reasoning ; it seems quite logical and convincing from the overall geological point of view.

As, however, our possibilities of understanding the meteorological conditions during ancient geological periods are very limited, we are accustomed to think that basic meteorological relationships which are valid today have also to be accepted for the ice-ages. We also customarily agree that glacial-meteorological conditions valid during the period immediately following the last ice-age can be accepted for drawing conclusions about more ancient geological periods. It is in relation to what we know about conditions on actual glaciers, and about the period of climatic optimum after the Würm ice-age, that I should like to draw attention to the following disagreements with the author's theory.

According to the author the disappearance of the last Würm-ice was due to a decrease of the radiation from the sun. This is turn is supposed to have been accompanied by a certain decrease of temperature but above all by a considerable decrease of precipitation, which is considered to have caused such unfavourable accumulation conditions that the ice-masses disappeared.

According to what we know from geological and vegetational evidence about the period during which the ice disappeared there is no evidence that the disappearance was simultaneous with a temperature decrease. As to the precipitation there were intervals both with a comparatively dry climate and with a rather wet climate. No indication exists of any overall trend towards a drying up in the areas of the earth where the snow accumulation conditions are of importance for the life of glaciers and ice-caps.

Turning to what we know of the regime of actual glaciers it has been rather clearly shown that the recent diminishing of the world's glaciers has been caused by an increase of summer temperature and humidity and also by a prolongation of the ablation season due to an increase of the winter temperature. Everything indicates that a negative regimen of glaciers generally is maintained by favourable radiation and convection conditions whilst changes in the accumulation conditions are of secondary importance. In order to get a considerable extension of glaciers and ice-caps, however, the most favourable condition seems to be an increase of winter precipitation; in other words ablation conditions play the dominant role in periods of diminishing of ice-masses, accumulation conditions are more important in the reverse process.

These results seem not to agree with the theory of the author that inter-glacial periods may have been connected with a lowering of both temperature and precipitation. Facts seem to make it particularly unlikely that the actual interglacial should be of this type.

Prof. H. C. Willett (Massachusetts Institute of Technology) : I find this paper of Sir George Simpson's to be perhaps the most intriguing and illuminating study of the problem of the glacial-interglacial climatic sequence that has appeared. In it the author demonstrates at its best his remarkable talent for cutting through obscuring complexities to get at the basic principles that underlie any problem in a commonsense, clear and concise manner, to be able to see the forest in spite of the trees. Whether one finds Simpson's basic thesis acceptable or entirely unacceptable, one cannot but admire the clear and simple logic by which he fits the pieces of his puzzle together.

I happen to be one of those who have long found Simpson's basic thesis of the solar radiation theory of glaciation essentially acceptable. However, I am not entirely convinced that the complete story, even accepting the solar hypothesis of glaciation, is nearly as simple as Simpson makes it appear, although it may possibly be so. There seem to be two points at which serious question may most obviously be raised in connexion with Simpson's hypothesis. The less important, perhaps, is the question of the true variability of the solar constant. Is there any indication that the over-all solar constant may be variable to the degree that Simpson would require ? Certainly there is no indication in the comparatively short record of Smithsonian solar constant determinations that this quantity is significantly variable today. On the other hand, this record is very short, serious doubt has been raised as to whether this technique accurately measures what it is supposed to, and the astronomers apparently are not averse to the idea that the sun may be a moderately,

though not strongly, variable star. There is the question of the quantitative determination of the solar corpuscular and ultraviolet radiations, which the astronomers agree certainly are very large percentually, but the absolute amount of which remains completely unmeasured by any direct means. The entire problem of the spectral distribution and amount of the variability of the solar output of energy remains to be determined by improved techniques of measurement.

The more important objection may be raised to Simpson's hypothesis that the observed climatic warming of the past half century has been accompanied by significant recession, not advance of glaciation as Simpson's hypothesis would require at this stage. However, it is to be noted that this temperature amelioration has been almost exclusively a phenomenon of the higher latitudes of the northern hemisphere, hence cannot be accounted for by any general increase of the solar constant. It has acted to decrease the poleward temperature gradient, hence to weaken the general circulation of the hemisphere. By Simpson's hypothesis this change should act strongly to decrease glaciation, as it has been observed to do. There is some indication that glaciation has increased moderately in recent decades in the high latitudes of the southern hemisphere. It might be remarked in closing that strongly increased solar corpuscular radiation, accompanying the rapid rise of sunspot activity in recent decades, logically may be considered as at least partly responsible for the pattern of the recent rise of temperature in the higher latitudes of the northern as opposed to the southern hemisphere.

Dr. K. P. Oakley (British Museum (Natural History)) : That Sir George Simpson has further developed his theory of Quaternary climatic change and used it to deduce the details of the relationship between recorded glacial and pluvial phases is immensely stimulating to all of us interested in Pleistocene correlations.

I should perhaps explain that when in 1949 I prepared the chart for *Man the Tool-maker* and decided to use the spread of early tool-traditions as approximate time-lines, I did so on the assumption that the time-lags involved were geologically negligible. That is to say the tradition of a new technique or fashion in tool-making might take 5,000 years to spread across a continent (actually I believe that would be an excessive estimate), but to regard a deposit of say 52,000 years old as " contemporaneous " with one 57,000 years old would generally be considered to involve only an insignificant error as far as Pleistocene geology is concerned. But is it negligible if we are concerned to test the validity of Sir George's thesis ? Climatic change can occur very rapidly, and it seems to me that until some of the critical deposit can be equated to within say 500 years it may be difficult to prove or disprove the relationships between pluvial and glacial phases which he infers. Minor climatic fluctuations are the cause of much confusion.

However, before long it should be possible to compare the radiocarbon dating of Kanjeran-Gamblian pluvial deposits with that of the main climatic fluctuations of the Last Interglacial and Last Glaciation in the Northern Hemisphere, and these results will perhaps bear conclusively on Sir George's postulates. A beginning has been made in America. Professor R. F. Flint recently referred in general terms to the dating of a pluvial freshwater layer in the Pleistocene evaporites beneath the dry floor of the Searles Lake in the arid part of south-eastern California. He says that radiocarbon dates *within* this pluvial layer and near the base of the overlying evaporites agree closely with dates of the classical Wisconsin (-Würm) drift of north-eastern United States of America, thus confirming the contemporaneity of glacial and pluvial conditions. The precise positions of the dated samples in the sequence are of critical importance. Thus, on the basis of Sir George's reconstruction of the pattern of change, only the *upper* part of the pluvial layer would be expected to equate with the maximum of the Wisconsin glaciation, while the overlying evaporite should equate with the later stages of that glaciation. The publication of the details of the investigation is therefore eagerly awaited.

It is most important that we should have agreed definitions of the terms glacial and pluvial. In theory there is much to be said for adopting Sir George's definition of a glacial stage as a time when the margins of the North polar ice-cap extended on the average south of lat. 75°N., but in practice this might be difficult to apply. His sub-division of pluvials into wet and very wet phases, and of interpluvials into very dry and less dry phases, is certainly worth testing in the field.

Geologists will be impressed by the author's authoritative statement as a climatologist that " if evidence of extensive ice action at sea-level is found at a locality between lat. 55°N and lat. 55°S, the land must have changed its latitude since it was glaciated." In Australia, extensive glacial deposits formed at or near sea-level during the Lower Permian age occur north of lat. 34°S. The only conceivable explanation is therefore continental drift or polar wandering.

Dr. F. E. Zeuner (Institute of Archaeology, London) : Sir George Simpson's admirable attempt to assess the effects of changes in solar radiation in terms of glacial and pluvial phases has to be viewed in the light of the geological evidence. The ultimate test of all theories of this

kind is, after all, the measure of agreement with the geological evidence. In this respect, I have the impression that his scheme is at variance with the observed facts. In Central Europe the study of loesses and buried soils has, in recent years, led to a generally accepted sequence of cool and temperate phases which cover the time from the Last Interglacial onwards. The Last Interglacial itself was, for a while, somewhat warmer than the Postglacial climate. It was followed by a cold phase (my LGl 1), this by a fully temperate phase comparable with the Postglacial, and this by the Last Glaciation in a restricted sense, divided into two cold oscillations by a mild period which never reached Postglacial temperate conditions. This rather complicated sequence of events is not reproduced in Sir George Simpson's diagrams.

Regarding the Mediterranean, it has been shown that the pluvials there are in close agreement with the fluctuations known from central and north Europe. About the tropical zone, however, which for meteorological reasons alone is likely to have had a different regime, little can as yet be said, since the climatic fluctuations that are believed to have existed there are in part based on the fillings of lake basins that experienced tectonic movements. Their levels, therefore, are not necessarily a function of climatic fluctuations.

In the southern hemisphere the most important sequence established is that of Söhnge, Visser and Van Riet Lowe which was based on the Vaal River gravels. This again is considerably more complicated than Sir George's scheme and I have not succeeded in reconciling it with the latter. The geological evidence for the fluctuations has been published in a large number of papers and summarized by me in *Dating the Past* (Methuen 1952). It appears that we are not yet in a position to make definite statements everywhere about the effects of solar radiation on the climate, at any rate on the warmer parts of the earth's surface, unless we restrict ourselves to the realm of pure hypothesis. In fact the evidence is still too ambiguous and has often been misinterpreted. Where detailed work has been done, the sequences established appear to be more complicated than those postulated by Sir George.

Mr. C. D. Ovey (Dept. of Geography, University of Cambridge) : The study of material collected from deep-sea cores suggests two important pieces of evidence which would not contradict Sir George's theory (Emiliani, C., *Journ. Geol.*, 1955, **63**, p. 538; Ericson, D. B. and Wollin, G., *Deep-sea Res.*, 1956, **3**, p. 104 and 1956, *Micropaleont.*, 1956, **2**, p. 257; Ovey, C. D., *Cent. Proc. R. Met. Soc.*, 1950, p. 211; Wiseman, J. D. H., *Proc. Roy. Soc.* A, **222**, p. 296.). First, from evidence provided by biological examination of the accumulated shell-remains of planktonic foraminifera, the rate of productivity of these shells (highest in warmer water), from CO_2 determination and oxygen and carbon isotope estimates of temperature and time, the recurrence of warm and cold periods of glacial and post-glacial times correspond happily with terrestrial evidence. Secondly, the cores so far examined have been taken from tropical and subtropical areas of the Atlantic. If these results correspond to northern-hemisphere warm and cold oscillations it seems reasonable to suppose that synchronous changes could have taken place in the southern hemisphere. Further work on deep-sea cores is very necessary but the practical difficulties in obtaining cores from temperate latitudes in the southern hemisphere are obvious.

One other point might be mentioned from the oxygen isotope studies of Emiliani. According to his estimates the temperature of the surface waters in the Caribbean and Equatorial Atlantic has varied between warm and cold periods between about 20°C and 30°C which indicates a very considerable change of temperature in the tropics between cold and warm periods. This would suggest too, that there was very much less available energy during the cold periods to provide the storminess necessary for Sir George's theory, although, in spite of the lower temperature, it might well be assumed that the temperature gradient between the edge of the ice-caps and tropical waters was much greater than it is today during our present " interglacial."

Sir George Simpson, F.R.S. (*in reply*) : I am very grateful to the meteorologists and geologists who have responded to the invitation of the Editor to express their views on my paper.

It is very encouraging that not one has questioned the fundamental theory that an increase in solar radiation above its present value would result in an enlargement of the polar ice-caps. The whole plan of my paper is to demonstrate that given the glacial epochs of high latitudes, the pluvial epochs of low latitudes are determined by the solar-radiation theory. It is therefore of the greatest importance that Dr. Leakey is able to write that the ideas set out in my papers seem to fit in very well with the recent work on the pluvial problem in East Africa.

It is clear, however, that I have not yet convinced all my commentators and most of them point out some difficulty in reconciling certain observations with the theory. This is not the place to discuss these difficulties individually. I shall, therefore, not consider each reply in detail, but will confine my remarks to general principles.

The first point I wish to make is that in the theoretical part of the work the only climatic factors I introduce are generalized glacial and pluval periods; therefore my theoretical conclusions can contain no information about any climatic change which is less than a whole glacial or pluvial period. Thus the climatic curve in Fig. 3 is not a complete account of all the small climatic changes which have occured during the Pleistocene; there may have been many short-period variations of the solar constant which are not represented on the curve. The small peaks on the pluvial record described by Dr. Leakey may be due to such small changes in the solar radiation.

On the other hand there are climatic changes which have nothing to do with changes in solar radiation and therefore should not enter into this discussion. Climatic changes of this nature can be recognized by the fact that they are not world-wide as they would be if they were due to changes in solar radiation. This is well illustrated by an example taken from the comments of Dr. Wallén and Dr. Willett.

It is now generally recognised that during the last half-century there has been a significant recession of the north polar ice-cap which has been accompanied by a definite warming of the region. On first sight this appears to be contrary to the theory. Both Dr. Wallén and Dr. Willett remark on this apparent discrepancy. Dr. Wallén finds that ' the facts do not agree with the theory ' but Dr. Willett sees that ' this temperature amelioration has been exclusively a phenomenon of the higher latitudes of the northern hemisphere, hence cannot be accounted for by any general increase of the solar constant.' Dr. Willett therefore has no fault to find with the theory on this account.

I do not think that it is generally realized how slow were the climatic changes which produced the Pleistocene sequence of glacial and pluvial phases. In the appendix to my paper I have made an approximate estimate of the difference of mean annual temperature at the equator (where the temperature changes are the greatest) between the maxima and minima of the climate curve in Fig. 3 and found it to be less than 18°C. Now the wavelength of the climate curve is 380,000 years on Oakley's scale, giving 200,000 years as the approximate time between the maximum and minimum temperatures; that is an average rate of change of 0·1°C in 1,000 years. Now this is a rate of change so small that it is entirely beyond the reach of instrumental measurement and its effect on climate, vegetation or animal life would be unrecognizable in the whole of historical time.

Dr. Oakley asks whether 5,000 years ' is negligible if we are concerned to test the validity of Sir George's thesis ' and adds ' climate change can occur very rapidly.' As I have already stressed in this reply, my thesis excludes rapid climatic changes which may or may not be due to variations in solar radiation. The temperature changes with which I deal are so slow that a period of 5,000 years would involve at the most a change of temperature of $\frac{1}{2}$°C; which obviously would be quite negligible in a test of my theory.

It is quite clear that there is much more work to be done on this interesting problem and I am continuing my studies in the hope of being able to put more numerical values of the temperature into the climatic record and to reach agreement with the geologists in a chronological table arranged according to the theory and including the results of their recent field work, especially in the British Isles. I should therefore welcome further criticisms and suggestions from meteorologists and geologists.

551.557.31 : 551.511.6

Theoretical studies of diurnal wind-structure variations in the planetary boundary layer

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SUMMARY

Averages of the wind vector at four times of the day were evaluated for 29 suitable days during the summer of 1951 at Wichita, Kansas, and Oklahoma City. The periodic portion of the wind variation vector is an ellipse at all levels with the major axis approximately in the direction of the wind vector. The amplitude reaches a maximum at 2,500 ft above the ground.

The cause of the variation is sought by theoretical methods, first by finding what periodic variations would occur when the eddy viscosity is periodic in time and constant with height, and secondly by finding solutions with the aid of an electronic analogue computer for cases when the eddy viscosity is distributed arbitrarily with height. It is concluded that variations of the type which were observed can occur only when both the average value of the eddy viscosity and the amplitude of its variations decrease rapidly with height above the lowest third of the friction layer.

1. INTRODUCTION

Fig. 1 is a composite diagram showing the average of the diurnal wind variations at different levels above the ground at Wichita, Kansas and Oklahoma City.

It was obtained from an analysis of routine 6-hourly pilot-balloon observations made on 29 days during the summer of 1951 which were selected so as to avoid marked changes of surface-pressure gradient. Owing to the fact that the direction of the pressure gradient differs from one day to another, the orientation of each wind was measured relative to the direction at 7,000 ft above sea level, at which level previous studies had indicated that

Figure 1. Composite diagram of diurnal wind variations at Wichita and Oklahoma City.

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the diurnal variation nearly disappears. The re-oriented wind vectors were averaged level-by-level and time-by-time so as to show the characteristic direction and speed of the wind at each level and time.

The odd values of height above the ground correspond to whole thousands of feet above sea level, both stations being at nearly the same elevation. Inasmuch as the days were not chosen consecutively, there was a fairly large difference between the mean wind at 1500 CST when the beginning of each day was reckoned and the value 24 hr later. This difference is ascribed to gradual changes of the pressure gradient during the period and was removed by subtracting from each wind an appropriate linear variation. Thus Fig. 1 represents only the periodic portion of the mean variation. The shape, orientation, and phase of the elliptical variations are similar to variations observed by Wagner (1939) in a study of wind variations in the mid-western United States. The variation of the wind at each level is much more pronounced in the direction of mean motion than in the direction at right angles thereto. Hence variations of speed are more noticeable than are variations of wind direction.

It is the purpose of this study to find what distribution of eddy viscosity in time and height best accounts for these features of the wind structure.

Before proceeding further it is well to consider whether the wind structure can actually be explained on the basis of the distribution of eddy viscosity with time and height. Wagner (1939) attributed the wind variations in the mid-western United States to a combination of (1) a circulation between the dry region of the south-west United States and its surroundings, (2) a large-scale land-sea breeze circulation, and (3) a large-scale mountain-plain circulation. Bleeker and Andre (1951) attributed the diurnal variation of convergence, which is probably associated with these same wind variations, to a large-scale mountain-valley wind system between the Rocky Mountains on the west side, the Appalachian Mountains on the east, and the broad Mississippi basin in between. That such circulations with a period of 24 hr and an amplitude of the order of 10 kt can arise through gravity drainage or thermal gradients seems quite implausible because of the tremendous mass of air involved in the circulations.

The most decisive argument against drainage as a cause of the observed variations concerns the phase of the variations. Heating and cooling of the lowest layers above a sloping surface result in a downhill force at night (in this case toward the east) and an uphill (westward) force during the day. Consider a coordinate system which is oriented so as to place the x-axis toward the east. The motions which result from a periodic force of the kind just described are given by the following equations, provided effects of advection and friction are ignored.

$$\left.\begin{aligned} \frac{\partial u}{\partial t} &= lv - \frac{1}{\rho}\frac{\partial \bar{p}}{\partial x} - F\cos\omega t \\ \frac{\partial v}{\partial t} &= -lu - \frac{1}{\rho}\frac{\partial \bar{p}}{\partial y} \end{aligned}\right\} \quad . \quad . \quad . \quad (1)$$

In these equations u and v are the velocity components in the x and y directions, l the Coriolis parameter, F the amplitude of the force, ρ the density, $\bar{p}$ the mean pressure over a 24 hr period, ω the angular velocity of the earth, and t the time after maximum surface temperature. It is sufficient for the purpose at hand to consider the pressure gradient as a constant. The solution may then be expressed by the following equations :

$$\left.\begin{aligned} u &= u_g + \omega F\,(l^2 - \omega^2)^{-1}\sin\omega t \\ v &= v_g + l\,F\,(l^2 - \omega^2)^{-1}\cos\omega t \end{aligned}\right\} \quad . \quad . \quad . \quad (2)$$

in which u_g and v_g are the components of the mean geostrophic wind. According to these equations the wind vector moves around an ellipse with north-south major axis when the

latitude exceeds 30°, and it reaches its maximum component towards the north in mid-afternoon. The mean wind direction at Wichita and Oklahoma City at 7,000 ft in summer is from the south-west. When the 7,000-ft wind direction in Fig. 1 is so oriented, it is seen that at lower levels the wind at mid-afternoon (1500 CST) actually has a minimum northward component. Therefore drainage and buoyancy effects do not offer a satisfactory explanation of the observed wind variations.

A theory of the diurnal variation of the wind has been given by Iswekow (1929) for eddy viscosity which is independent of height and varies sinusoidally with time. A solution was found by postulating that the ageostrophic wind is capable of representation as the product of a function of height and a function of time. The special solution thus found does not satisfy reasonable boundary conditions and is capable of satisfactory representation of the wind only in a restricted layer. In the next section this same problem is re-examined and a general solution is obtained which better explains several aspects of the observed variations.

Staley (1956) has criticized the use of sinusoidally varying exchange coefficient by several investigators and has found that simpler solutions based on a coefficient which is independent of time give equally good predictions of the height-time distributions of temperature changes. This simplification is unsatisfactory for a discussion of wind variations because, contrary to what is observed, it would result in a deviation from equilibrium which decreases with height. It is realized that the diurnal variation of exchange coefficient is in reality not exactly sinusoidal, but evidence to be presented herein indicates that the complications probably have little effect upon the wind variations above a few hundred metres. Thus no precise description of the form of the time variation should be expected from comparisons between theory and observations.

The procedure employed in this study is to find general solutions of the equations of motion for various distributions of eddy viscosity in time and height. Comparison with the observed variations then indicates the nature of the distributions of eddy viscosity which actually occurred. The geostrophic departure method, which was used by Lettau (1950) and by Sheppard and Omar (1952), was not attempted because a determination of the vertical distribution of the geostrophic wind was not feasible for so many days.

2. Analytical solution

The purpose of this section is to determine what type of wind variation occurs if the eddy viscosity is a periodic function of time and is independent of height. This analysis is not sufficiently powerful to be capable of explaining the observed variation at Wichita and Oklahoma City, but it serves a useful purpose in pointing the way for the possible solution of more difficult problems, and also in demonstrating the accuracy obtainable by an approximate method using a differential analyzer.

The wind variations which are associated with the diurnal and semi-diurnal variations of the pressure are much smaller than the variations observed in the boundary layer, and they are not limited to the surface layers (Bemmelen 1921, Johnson 1955). It is reasonable therefore to neglect time variations of the geostrophic wind. It will also be assumed that the geostrophic wind and the density are independent of height, that the mean wind vector is a function of height and time only, and that the vertical velocity is zero. The equations of motion can then be written

$$\left.\begin{aligned} \frac{\partial u^*}{\partial t} &= lv^* + \frac{\partial}{\partial z}\left[K(z,t)\frac{\partial u^*}{\partial z}\right] \\ \frac{\partial v^*}{\partial t} &= -lu^* + \frac{\partial}{\partial z}\left[K(z,t)\frac{\partial v^*}{\partial z}\right] \end{aligned}\right\} \quad . \quad . \quad . \quad (3)$$

where u^* and v^* are the components of the deviation of the mean wind from the mean geostrophic wind, l is the Coriolis parameter, and $K(z, t)$ is the kinematic eddy viscosity coefficient. In order to express these equations in a form that is convenient for solution, let

$$W = u^* + iv^* \qquad (4)$$

Let the complex number θ be defined by

$$\theta(z, t) = e^{ilt} W(z, t) \qquad (5)$$

Then the differential Eq. (3) may be replaced by

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K(z, t)\frac{\partial \theta}{\partial z}\right] \qquad (6)$$

The simplest case of diurnal variations occurs when $K(z, t)$, as given by

$$K(z, t) = k(1 + \eta \cos \omega t) \qquad (7)$$

in which η is a number between 0 and 1, ω is the angular velocity of the earth, k is a constant and t is measured from the time of maximum eddy viscosity. To find the solution for this case it is convenient to introduce the variables

$$\tau = t + \frac{\eta}{\omega}\sin \omega t \qquad (8)$$

and

$$\zeta = (l/2\pi^2 k)^{\frac{1}{2}} z \qquad (9)$$

Heights expressed by ζ are fractions of the gradient-wind height appropriate to k. For this case the differential equation reduces to

$$\frac{\partial \theta}{\partial \tau} = \frac{l}{2\pi^2}\frac{\partial^2 \theta}{\partial \zeta^2} \qquad (10)$$

It has been shown by Wagner (1936) that the daytime maximum of wind speed that occurs below a height of a few tens of metres is caused by the increase of eddy viscosity with height. The distribution of eddy viscosity assumed herein and the wind distribution which results from it can hardly be expected to apply to these layers. Therefore the lower boundary ($\zeta = 0$) is put near the top of this layer. It will be assumed that the variation of wind vanishes at this level and that it either vanishes or remains finite at infinity. After the resulting solution is found, the error resulting from actual variations at the lower boundary will be calculated and discussed. Accordingly, the following boundary conditions are adopted :

$$\left.\begin{aligned} &W(\tau, 0) = \theta(\tau, 0)\, e^{-ilt} = W_0 \\ &\lim_{\zeta \to \infty} \theta(\zeta, \tau) = \text{finite number} \end{aligned}\right\} \qquad (11)$$

in which W_0 is a constant.

Initial conditions need not be specified because only periodic solutions of the form

$$\theta(\tau, \zeta) = W_0\, e^{il\tau} \sum_{-\infty}^{+\infty} b_n(\zeta)\, e^{in\omega\tau} \qquad (12)$$

are desired. To find what functions must be used for $b_n(\zeta)$, one must substitute the periodic solution into the differential equation. After equating the coefficients of the various independent time functions, one obtains a set of differential equations

$$\partial^2 b_n/\partial \zeta^2 = 2i\pi^2(1 + n\omega/l)\, b_n \qquad (13)$$

each of which has an exponential solution. For n equal to 0, 1 and -1 respectively,

the solutions which satisfy the boundary condition at infinity are

$$\left.\begin{aligned} b_0(\zeta) &= b_0(0)\exp\left[-(1+i)\,\pi\zeta\right] \\ b_1(\zeta) &= b_1(0)\exp\left[-(1+i)\,(1+\omega/l)^{\frac{1}{2}}\,\pi\zeta\right] \\ b_{-1}(\zeta) &= b_{-1}(0)\exp\left[-(1+i)\,|1-\omega/l|^{\frac{1}{2}}\,\pi\zeta\right] \end{aligned}\right\} \quad . \quad . \quad (14)$$

In order to get a first approximation to the complete solution, it will be assumed that the coefficients of the higher-order terms vanish, and that the powers of η higher than the first can be neglected. In order to satisfy the boundary condition at $\zeta = 0$, the following equation must be valid.

$$e^{il(\tau-t)}\left[b_0(0) + b_1(0)\,e^{i\omega\tau} + b_{-1}(0)\,e^{-i\omega\tau}\right] = 1 \quad . \quad . \quad (15)$$

The inverse of Eq. (8) may be expressed as a power series in η as follows :

$$\left.\begin{aligned} \omega t &= \omega\tau - \eta\sin\omega\tau + \eta^2\cos\omega\tau\sin\omega\tau \\ &\quad - \eta^3\left(\cos^2\omega\tau\sin\omega\tau - \tfrac{1}{2}\,|\sin^3\omega\tau\right) + \ldots \end{aligned}\right\} \quad . \quad . \quad (16)$$

Since powers of η higher than the first are neglected, one obtains the expression

$$e^{il(\tau-t)} \simeq 1 + \frac{il\eta}{\omega}\sin\omega\tau = 1 + \frac{\eta l}{2\omega}\left(e^{i\omega\tau} - e^{-i\omega\tau}\right) \quad . \quad . \quad (17)$$

Substitution of this equation into (15) and requiring that the coefficient of each independent function of time be zero leads to the result :

$$b_0(0) = 1; \quad b_1(0) = -l\eta/2\omega; \quad b_{-1}(0) = +l\eta/2\omega \quad . \quad . \quad (18)$$

The final solution for the case $\phi \geqslant 30°$, is

$$\left.\begin{aligned} W(\zeta,\tau) &\simeq W_0\exp\left[-(1+i)\,\pi\zeta\right] \\ &+ (W_0\,\eta l/2\omega)\left\{\exp\left[-(1+i)\,\pi\zeta\right] - \exp\left[-(1+i)\,(1+\omega/l)^{\frac{1}{2}}\,\pi\zeta\right]\right\}e^{i\omega\tau} \\ &+ (W_0\,\eta l/2\omega)\left\{\exp\left[-(1+i)\,(1-\omega/l)^{\frac{1}{2}}\,\pi\zeta\right] - \exp\left[-(1+i)\,\pi\zeta\right]\right\}e^{-i\omega\tau} \end{aligned}\right\} \quad . \quad (19)$$

An example of this solution is plotted in Fig. 2 for latitude 37°N (the approximate latitude of Oklahoma City and Wichita). The wind at $\zeta = 0$, as computed from the theory described by Haurwitz (1941 p. 201), has been added in order to show the relation of the wind variations to the actual wind.

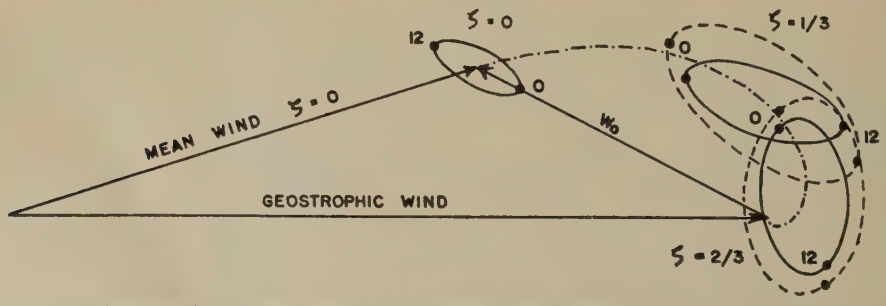

Figure 2. Approximate analytical solutions for eddy viscosity independent of height, $\phi = 37°$N. The velocity scale is arbitrary. Time is measured in hours after maximum eddy viscosity. Variations shown are for unit value of η and should be reduced in proportion to the value of η assumed. Dashed curve results from neglecting wind variations at $\zeta = 0$; solid curve takes such variations into account.

When η is zero there is no diurnal variation of eddy viscosity. According to Eq. (19) the wind vector is then distributed with height according to the Ekman spiral. When η is different from zero the wind vector executes an elliptical time variation, centred in the first approximation upon the equilibrium wind velocity. The amplitude of the time variation increases from zero at $\zeta = 0$ to a maximum and then decreases to zero again at great elevations; at any level, this amplitude is proportional, in the first approximation, to η.

Fig. 1 shows, as might have been expected from the work of Köppen (1921), that there is no level above the surface layer where the wind-vector variation actually vanishes. The error resulting from the erroneous boundary condition can be calculated in this case and will serve a useful purpose for estimating the error of more difficult subsequent calculations. Let the following boundary condition replace that in Eq. (11):

$$W(\tau, 0) = (1 + A\eta\, e^{i\omega\tau} + B\eta\, e^{-i\omega\tau})\, W_0 \quad . \quad . \quad . \quad (11a)$$

in which A and B are complex constants to be chosen so as to represent any desired elliptical variation at the lower boundary. Then the following correction should be added to the solution already found above.

$$\left.\begin{aligned} \Delta W = {} & A\eta\, W_0 \exp\left[-(1+i)(1+\omega/l)^{\frac{1}{2}}\pi\zeta\right] e^{i\omega\tau} \\ & + B\eta\, W_0 \exp\left[-(1+i)\left|1-\omega/l\right|^{\frac{1}{2}}\pi\zeta\right] e^{-i\omega\tau} \end{aligned}\right\} \quad . \quad . \quad . \quad (20)$$

This correction is a maximum at the lower boundary and decreases exponentially to zero at infinity, except at 30° latitude where its magnitude approaches a constant. An example showing the nature of this correction for a surface variation like that at the anemometer level in Fig. 1 is shown in Fig. 2. The correction has the effect of decreasing the amplitude and of rotating each of the ellipses slightly in a cyclonic sense.

Comparison of Figs. 1 and 2 indicates that η is probably not very small. To obtain better approximations when η is large, one must retain higher-order terms in Eq. (12) and correspondingly higher powers of η in Eq. (16). In this study the approximation was carried out to the third order, but because of its complexity the result is not repeated here. A plot of the third approximation of a representative case is shown in Fig. 3. These higher-order approximations result in considerable displacements of the points for large η, but the orientation and phase of the variations remain like those of the first approximation.

A fundamental discrepancy exists between the theoretical and observed distributions. The major axes of the elliptical variation in the theoretical case are inclined at a large angle to the wind direction, especially in the middle and upper portion of the boundary layer, whereas the actual ellipses have axes which are in all cases more nearly parallel to the direction of the wind at the same level. Thus the assumed distribution of eddy viscosity appears to be incapable of explaining the observed variations of the wind. Such an explanation requires a technique which allows other distributions of eddy viscosity with height. Also, in order to explain the magnitude of the variations which are observed, it is important to utilize a method which can deal with larger values of η.

3. Analogue method of solution

Even if some method can be found to remove the severe limitations of the analytical method of solution, there remains the difficulty that the computation and plotting of results are extremely laborious. A method is needed for finding quick solutions for any assumed distribution of eddy viscosity so that many different situations can be tried in a reasonable time. Such a method need not be exact. If the correct distribution of eddy viscosity can be indicated by an approximate result, it may later be feasible to verify the result by a more laborious method.

The Analogue Computer Laboratory of the Electrical Engineering Department of New York University has been of great assistance during these studies. The principal equipment of this laboratory is an electronic differential analyser, Model 16-31R, and two cathode-ray oscilloscopes which are connected directly to the analyser, one for visual observation, and the other for simultaneous photographic recording. The computer has a capacity of 20 summation units, 8 of which may also be used as integrators. It also has 4 multipliers, 32 coefficient potentiometers (which allow multiplication by a constant), and 6 output recorders.

In order to apply this type of analogue equipment to the problem, it is necessary to transform the two simultaneous partial differential Eq. (3) in two independent variables into a larger set of simultaneous ordinary differential equations in a single independent variable. The general method by which this is done is to replace the continuously distributed dependent variables u^* and v^* by a set of variable values at discrete levels. The differential equations for these discrete values are written by replacing the partial derivatives with respect to height by finite-difference approximations.

Let the index j refer to one particular level, $(j+1)$ and $(j-1)$ to the next higher and next lower level, respectively, and let a be the distance between levels. Then the simplest transformations of Eq. (3) are the following set of equations :

$$\left.\begin{aligned}\frac{du_j^*}{dt} &= lv_j^* + \left(\frac{K_j}{a^2} + \frac{K_j'}{2a}\right) u_{j+1}^* - \frac{2K_j}{a^2} u_j^* + \left(\frac{K_j}{a^2} - \frac{K_j'}{2a}\right) u_{j-1}^* \\ \frac{dv_j^*}{dt} &= -lu_j^* + \left(\frac{K_j}{a^2} + \frac{K_j'}{2a}\right) v_{j+1}^* - \frac{2K_j}{a^2} v_j^* + \left(\frac{K_j}{a^2} - \frac{K_j'}{2a}\right) v_{j-1}^*\end{aligned}\right\} \quad (21)$$

In these equations, K_j is the kinematic-exchange coefficient at the jth level, and K_j' is its vertical derivative at the same level. For purposes of obtaining a solution, these quantities are specified functions of time and are either generated by the computer, or, in some cases, fed into it from an external voltage source. The experimental procedure is to vary the K_j s and K_j' s to fit different trial patterns and to observe the solutions which result.

The number of levels which may be included in such a scheme is limited by the number of integrators in the computer. In this case, two integrators are required to generate a cosine function, leaving six to solve for dependent variables. Since the equations are of first order in time, six dependent variables can be accommodated. The model which is adopted consists of five levels numbered from 0 at the bottom to 4 at the top. At the bottom level the deviation from geostrophic wind is assumed to be constant, and the values

$$u_0^* = -100 \text{ and } v_0^* = 0$$

expressed in arbitrary units are adopted. At the top level, both components of the deviation are assumed to vanish. These conditions are similar to those which were imposed in the analytical problem and are acceptable if the height scaling is chosen so as to place the top level high enough to neglect the diurnal variation and if an allowance is made for non-negligible surface variations. When applied to the three intervening levels, the set of Eq. (21) gives six simultaneous ordinary differential equations to be solved for the three pairs of values of the wind components.

The choice of a suitable value for the height unit a is a compromise. The smaller a is allowed to be, the better is the finite-difference approximation of the vertical derivatives; however, a must be large enough so that the error involved in the upper-boundary condition is not objectionable. Trial comparisons between finite-difference calculations and analytical calculations indicate that a is best chosen so as to place the gradient-wind level appropriate

o k at the third level. This scaling is accomplished by the relation :

$$a^2 = \bar{k}/0{\cdot}456\, l$$

$\bar{k}$ being the average value of the kinematic eddy viscosity at the three levels.

TABLE 1. COMPARISON OF COMPUTER AND ALGEBRAIC SOLUTIONS OF THE DISCRETE-LEVEL MODEL AND THE EXACT SOLUTION (EKMAN SPIRAL) WHEN THE GRADIENT-WIND LEVEL IS LOCATED AT $z = 3a$

	a		$2a$		$3a$		$4a$	
	u^*	v^*	u^*	v^*	u^*	v^*	u^*	v^*
1) Algebraic	− 19·4	+ 26·4	+ 3·3	+ 10·4	+ 3·3	+ 1·5	0·0	0·0
2) Computer	− 19·0	+ 27·5	+ 3·6	+ 11·0	+ 3·5	+ 1·5	–	–
3) Theoretical	− 17·5	+ 30·3	+ 6·2	+ 10·7	+ 4·3	0·0	+ 0·8	− 1·3

The simplest estimate of the errors of this method is furnished by a comparison of three solutions of the wind distribution for constant eddy viscosity : (1) an exact algebraic solution of the finite-difference equations, (2) a computer solution of the same, and (3) an exact calculation of the Ekman-spiral solution for the same conditions. Comparison of (1) and (3) in Table 1 indicates that the use of the discrete-level model involves an error of about 2 per cent of the wind deviation at the surface. The error which the computer makes in solving the finite-difference equations, which is shown by a comparison of solutions (1) and (2), is less than 1 per cent of the surface deviation. A more realistic comparison can be made between the results of the analytical solution for identical parameters (see Fig. 3). The two solutions are displaced from each other by a nearly constant amount, and this displacement is about equal to that between the algebraic and theoretical solutions in Table 1. Its cause is the crudeness of the discrete-level model. The remaining errors are small and do not affect the amplitude, shape, or phase significantly, as may be judged by referring to Fig. 3, in which the computer solution has been shifted a constant amount

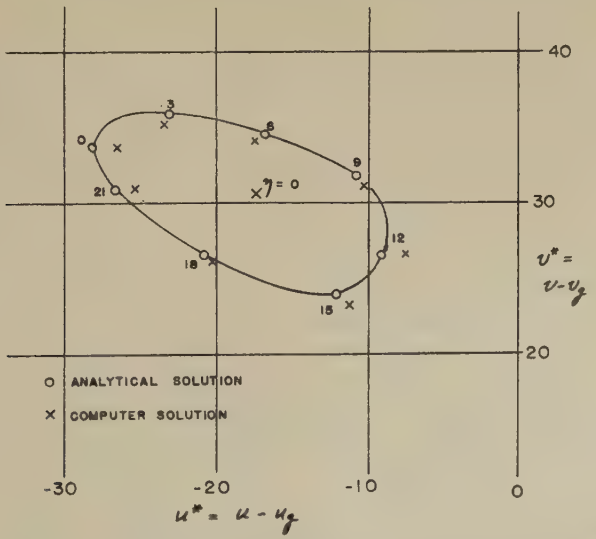

Figure 3. Comparison of the analytical solution (third approximation) and the modified computer solution for the case $K = k\,(1 + \eta \cos \omega t)$, $\zeta = \frac{1}{3}$, $\phi = 45°$, and $\eta = \frac{1}{3}$. The computer solution has been given a constant displacement so as to make the solutions agree when $\eta = 0$.

in order to bring the equilibrium solutions into agreement. The analogue method thus appears to be a very adequate tool for exploring the effect of various distributions of eddy viscosity in time and height upon the nature of the resulting variations of the wind.

Three varieties of functions were used for $K(z, t)$ during the course of the experiments, each of which was associated with a different programme for the computer. These functions are

$$\left.\begin{aligned} K_1(z,t) &= k(z)(1+\eta\cos\omega t) \quad (\eta = \text{const}) \\ K_2(z,t) &= k(z) + L(z)\cos\omega t \quad L(z) \leqslant k(z) \\ K_3(z,t) &= k(z) + L(z) f(t) \end{aligned}\right\} \quad . \quad . \quad (22)$$

in which $k(z)$, $L(z)$ and $f(t)$ are arbitrary functions. The function $K_1(z, t)$ is a special case of the functions $K_2(z, t)$ and $K_3(z, t)$ but is treated separately because it uses a simpler programme. The functions $K_2(z, t)$ could not be completely programmed because of limitation in the number of summation units; the programme which was adopted allows L to be given different values at each level, but requires $\partial L/\partial z$ at each level to be neglected. That the programmes for $K_2(z, t)$ and $K_3(z, t)$ nevertheless produce solutions of value is indicated by the fact that a repetition of some solutions of the K_1 programme using the K_2 programme gave similar though not identical results despite the fact that $\partial L/\partial z$ was actually different from zero.

4. Analogue studies of wind variations

Once the analogue computer has been programmed for the form of $K(z, t)$, it is only necessary to adjust the potentiometers for appropriate values of k, $\partial k/\partial z$, and either η or L at each of the levels. The machine is then allowed to run from arbitrary initial conditions through several cycles until inspection of the recorder indicates that the solution is completely periodic, usually within one minute. In a sense the computer functions as a model of the atmosphere which is amenable to controlled experiments. More than 100 experiments representing different sets of distributions of exchange coefficient in time and height were conducted. Only a few of the more interesting ones will be discussed here.

The graphs which are given in Figs. 4-10 represent in each case the deviations from the geostrophic wind in arbitrary units. The value of this deviation vector which has been assumed at the lowest level as boundary condition is also shown. In attempting to compare the computed wind variations with the observed values which are shown in Fig. 1, one should bear in mind the relation of the vector W_0 to the surface wind and the geostrophic wind. The conditions which have been assumed about the distribution of eddy viscosity during the conduct of the experiments do not specify this relation between W_0 and the geostrophic wind; it is determined instead by the distribution of eddy viscosity in the surface layers. Comparison of the ensuing figures with Fig. 1 therefore requires some inference about the wind at $z = 0$. Perhaps the vectors shown in Fig. 2 will serve this purpose.

Whenever an attempt is made to compare computer solutions with the variations of the wind shown in Fig. 1, it should be kept in mind that the times shown in Figs. 4 to 10 are hours after the time of maximum eddy viscosity, which may be presumed to occur at about the time of maximum surface temperature. In most cases the time 0 hr in the computer solution may be compared with the time 1500 CST in Fig. 1.

The distributions of eddy viscosity which were used in the experiments which are about to be described are summarized in Table 2.

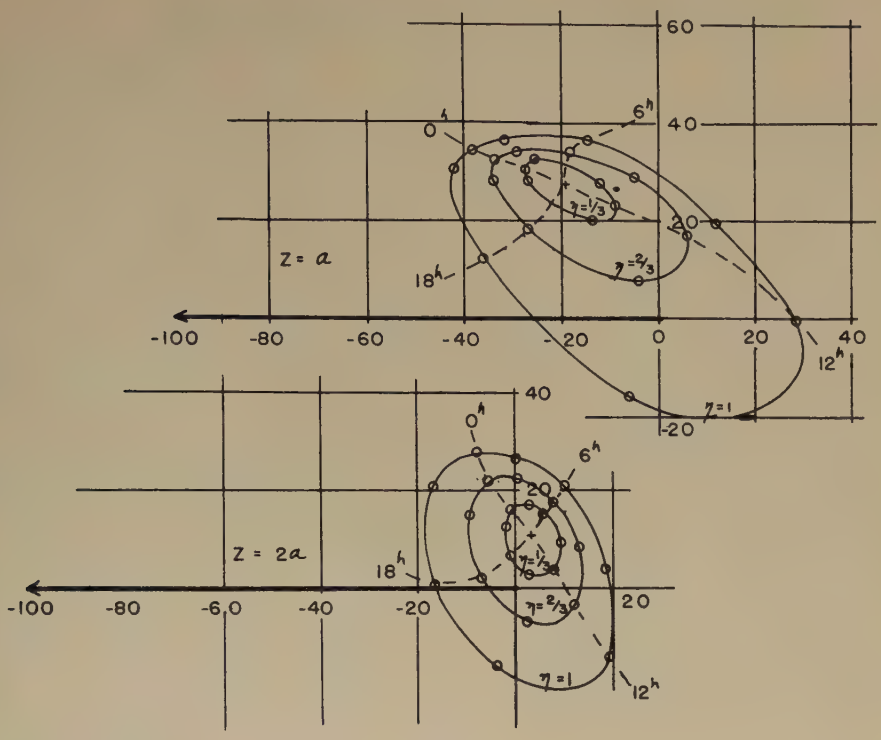

Figure 4. Wind variations at $z = a$ and $z = 2a$ for various values of η when $k(z) = 0{\cdot}456\ a^2 l$ at latitude 45°.

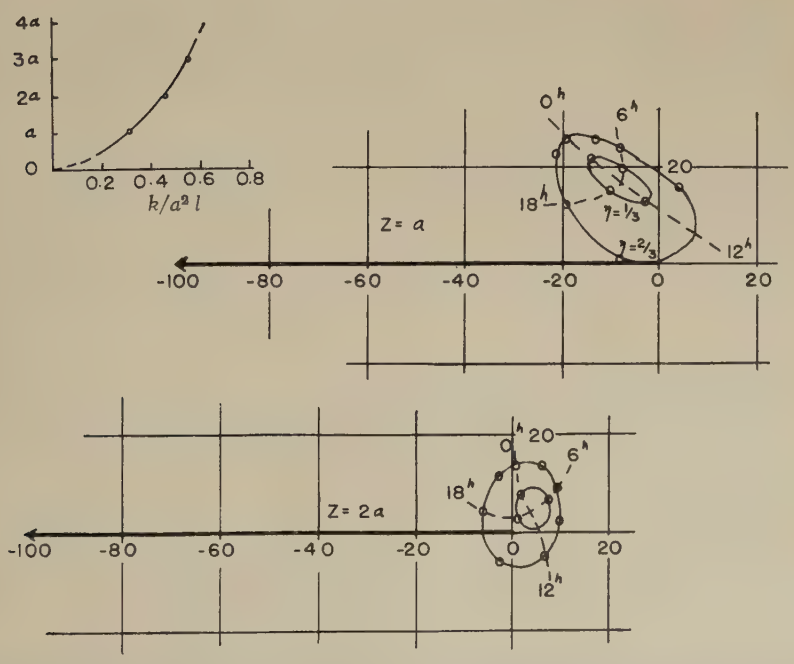

Figure 5. Wind variations at 45°N for values of $\eta = \frac{1}{3}$ (inner) and $\frac{2}{3}$ (outer) with mean eddy viscosity and amplitude of eddy viscosity both increasing with height as shown in the inset.

TABLE 2. PARAMETERS USED FOR EXPERIMENTS DESCRIBED IN THIS REPORT.
(In Fig. 8 the values given for L/k were not used).

Figure	$k_1/a^2 l$	$k_2/a^2 l$	$k_3/a^2 l$	$k_1'/a l$	$k_2'/a l$	$k_3'/a l$	L_1/k_1	L_2/k_2	L_3/k_3
4	0·456	0·456	0·456	0·000	0·000	0·000	–	–	–
5	0·322	0·456	0·558	0·162	0·114	0·094	–	–	–
6	0·370	0·400	0·450	0·060	0·030	0·090	0·230	0·160	0·050
7, 8, 10	1·000	0·400	0·100	− 0·500	− 0·225	− 0·075	0·800	0·800	0·800

A comparative study of Figs. 4 to 7 reveals a number of significant facts :

(1) The relation between the amplitude of the eddy viscosity and the amplitude of the wind variations is non-linear; large values of η bring disproportionately larger amplitudes than smaller values.

(2) With similar amplitudes of eddy viscosity, the largest amplitude oscillations of wind occur when both the mean value and the amplitude of the eddy viscosity decrease rapidly with height at and above the first level (such a distribution implies a lower-level maximum, since the eddy viscosity increases upward at the surface).

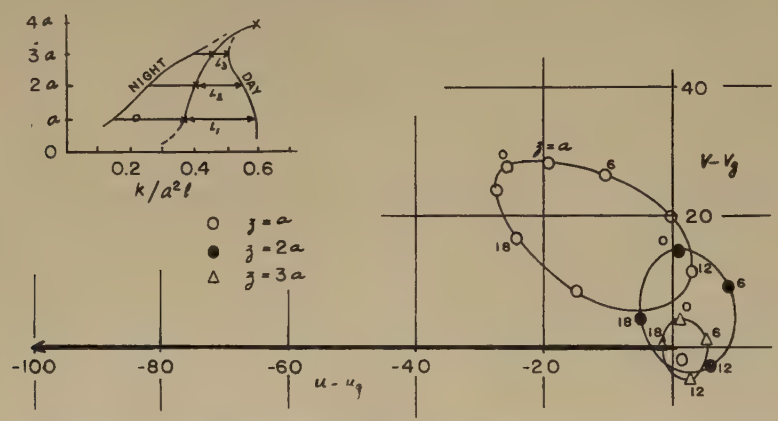

Figure 6. Wind variations at 45°N when the mean eddy viscosity increases with height and the amplitude of the variation of eddy viscosity decreases with height as shown in the inset.

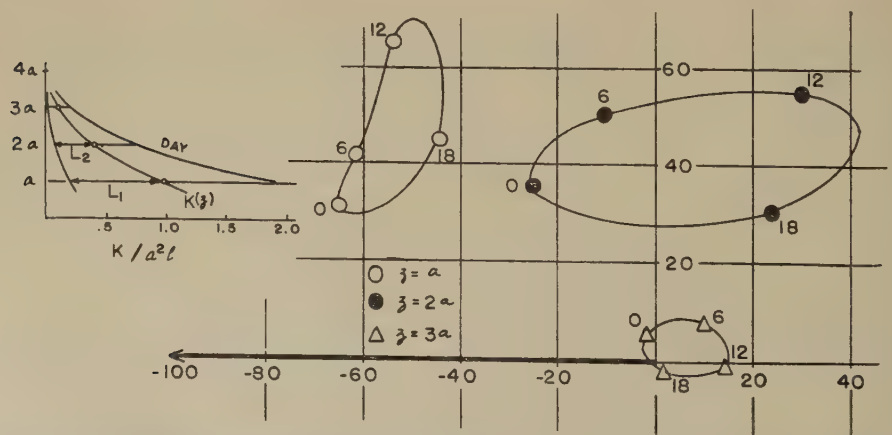

Figure 7. Wind variations at 45°N when both the mean eddy viscosity and the amplitude of the variation of eddy viscosity decrease with height at and above the first level, as shown in the inset diagram, using the computer programme for $K(z, t) = k(z) + L(z) \cos \omega t$.

(3) The orientation and phase of the elliptical variations are most nearly like those which are observed when both the mean eddy viscosity and the amplitude of its variation decrease rapidly with elevation.

(4) The largest of the computed amplitudes is not more than half as large as that of the observed oscillations in Fig. 1. It is recalled that if allowance is to be made for more realistic surface boundary conditions the computed amplitude of the oscillations must be even smaller. It thus appears that an even larger diurnal range of eddy viscosity than that used in Fig. 7 must actually occur.

(5) There is an inverse relation between the magnitude of the wind and the eddy viscosity. For example, the wind difference between the surface and the first level is smaller in Fig. 7, where the eddy viscosity is large, than in Fig. 5 where it is small. Between the second and third level in the same two figures the situation is reversed, with, however, the same relation between wind shear and eddy viscosity.

The amplitude, orientation and phase of the observed variations are best accounted for in Fig. 7 in which both the mean value and amplitude of the eddy viscosity decrease with height. Since the programme used in obtaining this result necessitated a neglect of the vertical derivative of the amplitude of eddy viscosity, another calculation, using a similar distribution which is described by giving η a constant value, is instructive. Such a case is shown in Fig. 8.

The mean position of the wind vector agrees very closely with that of Fig. 7, but the orientation of the ellipses is different, agreeing more nearly with that of the other experiments. The rather different orientation in Fig. 7 may therefore be due to neglect of the terms involving the vertical derivative of the amplitude, since the earlier examples did not indicate much change of orientation with variations of η. Although the results shown in Fig. 8 do not agree as well with the observed wind distribution as the results in Fig. 7, they are nevertheless much better in agreement than any of the others. For example, Fig. 8 shows also the variations at corresponding levels when $k(z)$ is constant with height and η is also 0·4. The result for k decreasing gives more nearly the observed orientation and a slightly better phase at both levels. The larger variation of amplitude of the wind at the second level shows better promise of being able, with larger values of η (which unfortunately were not attainable without overloading components of the computer), to explain the enormous amplitude of the observed wind variations. It may be remarked that because of its interest the experiment shown in Fig. 8 has since been recomputed using an analytical method which is being developed by K. Ooyama. The orientation, phase and amplitude agree very closely, but there is some discrepancy of the mean location of the ellipses, such as has been noted in the earlier comparisons between computer and analytical solutions.

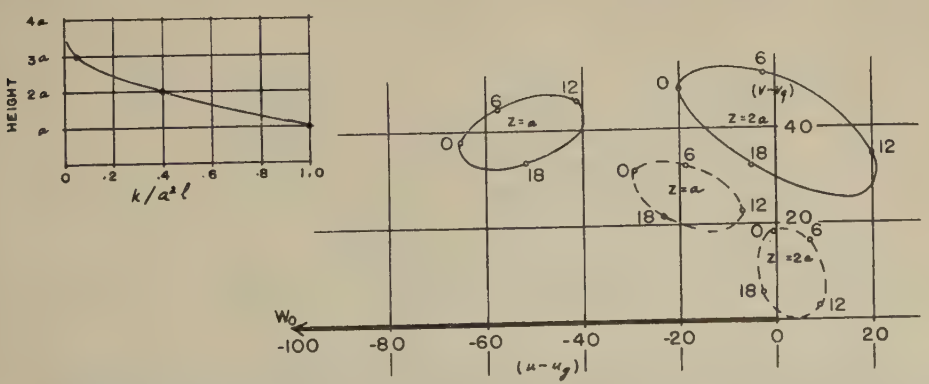

Figure 8. Solutions for a case when $k(z)$ decreases rapidly with height (as shown in the inset); $\eta = 0{\cdot}4$; $\phi = 45°$ (solid lines). Dashed curves are solutions for $k(z)$ constant, $\eta = 0{\cdot}4$, and $\phi = 45°$.

The behaviour of the wind variation with change of latitude is illustrated by an experiment shown in Fig. 9. The eddy-viscosity distribution is the same as that shown in Fig. 7 except that the range of the variation has been reduced slightly to prevent overloading of the computer at low latitudes. The experiment shows a typical decrease of of amplitude with increasing latitude between 30° and 55°N. The larger amplitude of 30° could be expected from the fact that the free period of inertia oscillations becomes equal at that latitude to the period of the forcing function. The tendency for resonance is actually not excessive because a large amount of viscous damping is characteristic of these layers.

It is of interest to investigate the sensitivity of the solutions to a change of the time function used to describe the diurnal variation of eddy viscosity. The inset of Fig. 10 shows one of several selections of function $f(t)$ for use in the third equation of (22); it is characterized by a relatively brief maximum and long minimum and in this respect approximates observed low-level variations of exchange coefficient more closely than a simple cosine curve. It was obtained by a single integration of a manually-generated

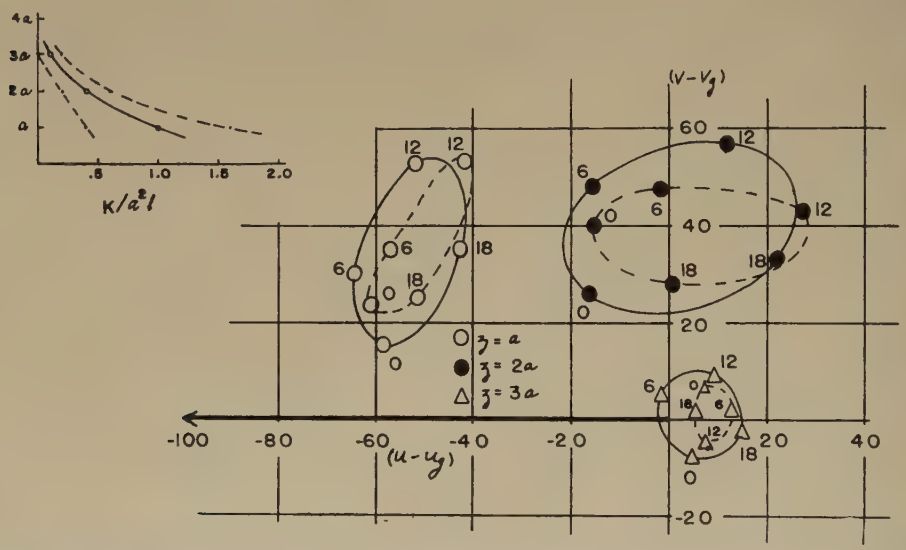

Figure 9. Wind variations at 30°N (solid curves) and at 55°N (dashed) when both the mean eddy viscosity and amplitude of the eddy viscosity decrease with elevation as shown in the inset.

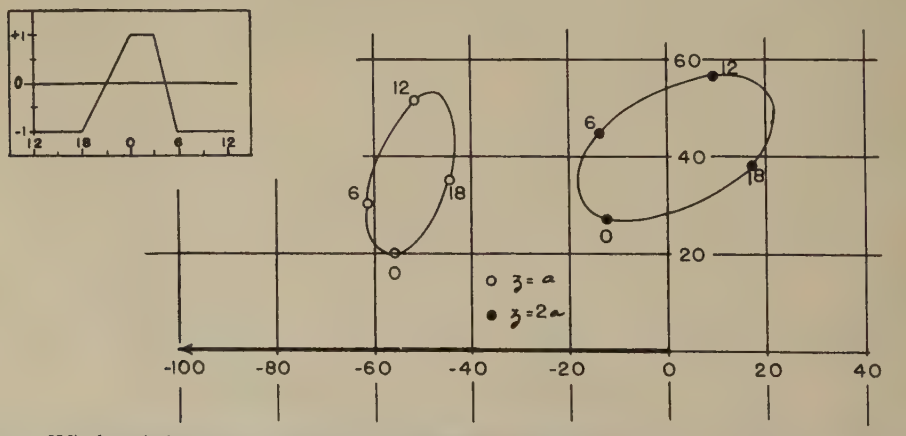

Figure 10. Wind variations at 45°N with a non-sinusoidal variation of eddy viscosity. Inset shows the function $f(t)$ in Eq. (22). The vertical distributions of the mean eddy viscosity and the amplitude are the same as in Fig. 7.

square wave. Except for this substitution of input the programme and constants used were the same as for the case shown in Fig. 7. The resulting wind variations are shown in Fig. 10. Some differences ascribed to the alteration are discernible, but the general conclusion reached from this and several other experiments is that the substitution of any periodic function bearing a vague similarity to the cosine function does not alter the wind variation in any important respect.

Natural variations are subject to disturbances due to non-periodic variations of stability, latitude, cloudiness and physical properties of underlying surface. A rapid change of physical conditions is followed by a period of transition from one kind of wind variation to the periodic variation appropriate to the altered conditions. If the transition period is very long, the form of the observed quasi-periodic variation may not be related so much to the existing physical condition as it is to some combination of conditions during the past history of the air mass. It is a simple matter to observe this period on the computer by starting the solution at an initial value which is removed from the final solution and observing how many oscillations occur before the deviation from the completely periodic solution becomes imperceptible. In most cases not involving extreme initial conditions the transient period lasts for only $\frac{1}{2}$ to $\frac{3}{4}$ period. Therefore, the variations which occur on a given day are not likely to display any perceptible effect of conditions existing during the previous day. An air mass which moves inland from the sea where there was no diurnal wind variation at all may acquire a fully developed frictional-type diurnal wind variation in as little as 12 hours.

5. Summary and conclusions

None of the more than 100 experiments which were carried out on the computer gives a satisfactory explanation of the orientation and amplitude of the observed diurnal wind-structure variation at Oklahoma City and Wichita. The best agreement occurs when both the mean value of the eddy viscosity and the amplitude of the variation decrease rapidly with height above the surface layers. The distribution of eddy viscosity with height probably depends upon the temperature structure of the air mass; therefore the distribution of eddy viscosity which appears to have occurred in connection with the wind variations shown in Fig. 1 is not necessarily typical of that which would be found in other air masses. The prevailing wind direction in this case was southerly. Brown (1955) found that diurnal variations of the wind are much smaller in amplitude when the prevailing wind direction is from the north, a difference which may be the result of a different temperature structure. The distribution which is indicated in this study is similar to that found by Mildner (1932) from an examination of the average of a large number of pilot-balloon observations made in Leipzig. He found that the dynamic-exchange coefficient reaches a maximum value of 500 g cm^{-1} sec^{-1} at the height 240 m and decreases rapidly to a value of 70 g cm^{-1} sec^{-1} at 500 m. A re-examination of the same profile by Lettau (1950) has yielded a qualitatively similar distribution of eddy viscosity with, however, a smaller value of the maximum and a smaller rate of decrease with height.

Even when η is 1, so that amplitude of eddy viscosity is the maximum, the amplitude of the computed wind variation is too small to account for the observed amplitude so long as the eddy viscosity increases with height or remains constant. When the eddy viscosity decreases rapidly with height it may be possible to account for the large observed variation if η exceeds the value 0·8. Thus it appears that the eddy viscosity during the day reaches a value which is considerably larger than ten times the night-time value. The large range thus indicated is in agreement with the range of the exchange coefficient for heat transfer which has been observed near the ground (Staley 1956). The smallness of the night-time

values of eddy viscosity helps to account for the large super-geostrophic winds observed in connection with the boundary layer wind maximum which occurs regularly in the mid-western United States (Blackadar 1957).

The mean wind direction at 2,500 ft above the surface in the mid-western United States in summer is from the south-west. During the daytime the wind is more retarded in the south than it is in the north (see Fig. 9). Therefore, with a tendency for decreasing wind downstream, a field of divergence tends to become established. At night (between 12 and 18 hr after the time of maximum eddy viscosity in Fig. 9) the situation is reversed and convergence tends to be set up. Thus diurnal changes in the field of convergence occur which are qualitatively similar to that observed by Bleeker and Andre (1951). The magnitude of the latitude effect by itself is at least an order of magnitude too small. It is probable, however, that differences in the nature of the underlying surface (as for instance must be present between the area around the Great Lakes and the semi-arid region of the southwest) may be responsible for large enough differences of amplitude to account for the observed fields of convergence.

Inasmuch as theoretical methods have not given a completely satisfactory account of the observed wind variations, further observations and analysis of the wind structure and its relation to the geostrophic wind in this area will be of great interest.

6. Acknowledgments

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The formation of large droplets in small cumulus

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Summary

Droplets of radius exceeding 30 μ are sometimes found in concentrations of order 100/*l* in non-precipitating cumulus less than 5,000 ft in depth and having, on average, liquid-water contents of only a few tenths of a gramme per m^3. It is shown that droplets of this size may be produced by condensation and coalescence on hygroscopic nuclei of mass greater than 10^{-11} g entering the bases of such shallow clouds, provided that these last for at least one hour. The published concentrations of giant salt nuclei seem only just sufficient to account for the observed numbers of large droplets, but simultaneous measurements of both are urgently required.

1. Introduction

An important problem in cloud physics at the present time concerns the initial stages of shower production by the coalescence mechanism. It has been shown by Ludlam (1951) that if droplets of radius 20-30 μ enter the base of a warm cloud sustained by a steady updraught of 1-3 m/sec, a shower may be released, if the cloud depth exceeds about $1\frac{1}{2}$ km, within about 20-30 min. The suggestion was first made by Woodcock (1950) that the larger droplets required to start the coalescence mechanism might be supplied by the giant nuclei found by him in the sub-cloud layer over the sea. Nuclei of mass exceeding 10^{-9} g, which are found in concentrations of order 1/litre, leave the sea surface as spray droplets of radius greater than 20 μ, and in convective weather may be carried up into the cloud without appreciable evaporation. They may therefore be effective for shower production in unstable maritime air in warm weather. But as the air moves inland, they may disappear by sedimentation, be removed by precipitation, or shrink by evaporation towards their equilibrium radii appropriate to the prevailing relative humidity. Determination of the giant-nucleus content of the air and calculation of the rate at which such particles may grow after entering the cloud are therefore very relevant to the problem of shower production over the continents.

Cloud-droplet samples obtained by Weickmann and aufm Kampe (1953), and by the Meteorological Research Flight (Durbin 1956) certainly indicate that droplets of radius greater than 30 μ are sometimes present in appreciable concentrations in quite small cumulus. For example over southern England, Durbin reports that cumulus less than 5,000 ft deep and having liquid-water contents of only about 0·5 g/m^3 contain droplets of 30 μ radius in concentrations of order 100/l; 50 μ droplets are found in concentrations of up to about 1/l (Murgatroyd, private communication). Rather similar concentrations of 30 μ droplets were found by Weickmann and aufm Kampe in fair-weather cumulus over New Jersey and Florida.

In this paper we attempt to show how these large droplets may arise in such shallow clouds within a reasonable time, and discuss whether their numbers may be accounted for in terms of the probable giant-nucleus content of the air. We consider, first of all, droplet growth on a population of condensation nuclei as they are lifted from cloud base, along a saturated adiabatic, in a steady updraught. The supersaturation and droplet radii are calculated as functions of time in terms of the updraught speed, the temperature, pressure, and the size and concentration of the nuclei. Similar calculations have been made by Howell (1949), but the present formulation of the problem, leading to two rather simple equations for the time variation of supersaturation and droplet radius,

appears much more convenient for computation. The population of condensation nuclei is assumed to contain a very small number of giant nuclei which enter cloud base as droplets of saturated salt solution, i.e., with about twice their dry radius. Their subsequent growth by condensation and coalescence is then calculated in terms of the computed variation of supersaturation and liquid-water content with height. After having reached a height of $1\frac{1}{2}$ km the updraught is arrested, and thereafter it is assumed that some of the droplets which grew upon giant nuclei in the ascending bubble are mixed downwards to lower levels in the cloud, growing by coalescence on their journey through cloudy air of rather low liquid-water content.

2. Calculation of the supersaturation and rate of droplet growth in a cumulus cloud

(i) *Calculation of supersaturation*

The supersaturation of an air mass can be expressed in terms of the difference between its dew point T_d and the actual temperature T of the air. If saturated air at temperature T and vapour pressure e_s is warmed to $T + dT$ then, by the Clausius-Clapeyron equation, the incremental increase in vapour pressure de_s required to keep it saturated is given by $\frac{de_s}{e_s} = \frac{\epsilon L}{RT^2}\,dT$, where L is the latent heat of vaporization of water, ϵ the specific gravity of water vapour relative to that of dry air, and $R = 2{\cdot}87 \times 10^6$ erg g^{-1} °K^{-1} the gas constant for one gramme of dry air. If now the air is still kept saturated while being warmed to a slightly higher temperature $T + dT_d$ and thereafter cooled to $T + dT$ *without condensation occurring*, the vapour pressure will now be in excess of saturation by an amount de such that the supersaturation $d\sigma$ may be written

$$d\sigma = \frac{de}{e_s} = \frac{\epsilon L}{RT^2}\,(dT_d - dT), \qquad . \quad . \quad . \quad (1)$$

this equation being valid only for small changes in supersaturation.

Now the change in temperature of the air following a vertical displacement dz is given by $dT \simeq \left(-\frac{g}{c_p}\,dz - \frac{L\,dx}{c_p}\right)$, $\frac{g}{c_p}$ being the dry adiabatic lapse rate, dx the change in humidity mixing ratio and c_p the specific heat of air at constant pressure.

Thus, $$d\sigma = \frac{\epsilon L}{RT^2}\left(dT_d + \frac{g}{c_p}\,dz + \frac{L\,dx}{c_p}\right) \qquad . \quad . \quad . \quad (1a)$$

We may also use the following relations to represent small changes occurring in *slightly* supersaturated air :

$$dx + dw = 0, \qquad . \quad . \quad . \quad . \quad (2)$$

where w is the liquid-water mixing ratio in grams per gram of dry air;

$$dx = \epsilon\, d\left(\frac{e_s}{p}\right) = \frac{\epsilon}{p^2}\,(p\,de_s - e_s\,dp), \qquad . \quad . \quad . \quad (3)$$

$$\frac{1}{e_s}\frac{de_s}{dT_d} = \frac{\epsilon L}{RT^2}, \qquad . \quad . \quad . \quad . \quad (4)$$

and $$\frac{dp}{dz} = -\frac{gp}{RT}, \qquad . \quad . \quad . \quad . \quad (5)$$

where p is the total pressure and g the acceleration due to gravity.

Substitution from (4) and (5) into (3) gives

$$dx = -dw = \frac{\epsilon}{p^2}\left(\frac{e_s\, gp}{RT}\, dz + \frac{\epsilon p\, L\, e_s}{RT^2}\, dT_d\right) \quad . \quad . \quad . \quad (6)$$

and finally, substitution for T_d from (6) into (1a) yields

$$d\sigma = \frac{\epsilon L}{RT^2}\left[g\left(\frac{1}{c_p} - \frac{T}{L\epsilon}\right) dz - \left(\frac{L}{c_p} + \frac{pRT^2}{L\epsilon^2\, e_s}\right) dw\right] \quad . \quad . \quad (7)$$

If we consider the liquid droplets being carried up in an updraught U, then Eq. (7) gives an expression for the time variation of the supersaturation σ, viz :

$$d\sigma/dt = AU - B\, dw/dt \quad . \quad . \quad . \quad . \quad . \quad (8)$$

where $$A = \left(\frac{\epsilon Lg}{c_p RT^2} - \frac{g}{RT}\right) = \frac{3{\cdot}42 \times 10^{-4}}{T}\left(2{\cdot}6\,\frac{L}{T} - 1\right),$$

L being measured in cal g^{-1}, and

$$B = \left(\frac{\epsilon L^2}{c_p RT^2} + \frac{p}{\epsilon e_s}\right) = \left(37{\cdot}8\,\frac{L^2}{T^2} + 1{\cdot}62\,\frac{p}{e_s}\right).$$

(ii) *Equations of droplet growth*

For a droplet growing upon a small hygroscopic nucleus in a supersaturated atmosphere the governing equations (see Howell (1949), Mason (1951)), can be written as follows :

$$r\frac{dr}{dt} = \frac{DM}{\rho \mathbf{R} T}[e - e'(r)], \quad . \quad . \quad . \quad . \quad . \quad . \quad (9)$$

$$e'(r) = e_s(T_r)\left[1 + \frac{2\gamma M}{\rho \mathbf{R} T r} - \frac{8{\cdot}6\, m}{Wr^3}\right], \quad . \quad . \quad . \quad (10)$$

$$r\frac{dr}{dt} = \frac{K}{L\rho}(T_r - T), \quad . \quad . \quad . \quad . \quad . \quad . \quad (11)$$

$$\frac{1}{e_s}\frac{de_s}{dT} = \frac{LM}{\mathbf{R}T^2}, \quad . \quad . \quad . \quad . \quad . \quad . \quad (12)$$

where e, $e'(r)$ are respectively the vapour pressure of the environment and that existing at the surface of the droplet, $e_s(T_r)$ is the saturation vapour pressure at the surface temperature T_r of the droplet; γ, ρ, L, M are respectively the surface tension, density, latent heat and molecular weight of water, D the diffusion coefficient of water vapour in air, K the thermal conductivity of air, m the mass of the nucleus having molecular weight W; $\mathbf{R}$ is the universal gas constant and T the absolute temperature. Integration of (12) over the temperature range T to T_r gives :

$$\ln\frac{e_s(T_r)}{e_s(T)} \simeq \frac{e_s(T_r) - e_s(T)}{e_s(T)} = \frac{LM}{\mathbf{R}}\frac{(T_r - T)}{T_r T} \simeq \frac{L^2 M\rho}{K\mathbf{R}T^2}\, r\frac{dr}{dt}. \quad . \quad (13)$$

Substitution for $e'(r)$ from (10) into (9) gives

$$\frac{e - X e_s(T_r)}{e_s(T)} = \frac{\rho \mathbf{R} T}{DM\, e_s(T)}\, r\frac{dr}{dt}, \quad . \quad . \quad . \quad . \quad (14)$$

where $$X = \left(1 + \frac{P}{r} - \frac{Q}{r^3}\right); \quad P = \frac{3{\cdot}2 \times 10^{-5}}{T}, \quad Q = \frac{8{\cdot}6\, m}{W}.$$

Adding (13) and (14), we have :

$$r\frac{dr}{dt}\left[\frac{L^2 M\rho}{K\mathbf{R}T^2}+\frac{\rho\,\mathbf{R}\,T}{DM\,e_s(T)}\right]=\sigma-\left(\frac{P}{r}-\frac{Q}{r^3}\right)\frac{e_s(T_r)}{e_s(T)}, \quad . \quad . \quad (15)$$

in which $e_s(T_r)/e_s(T)$ is very close to unity.

3. Growth of a droplet population

If we now consider n droplets, each of radius r, constituting w grams of liquid water per gram of dry air, to be lifted from cloud base, along a saturated adiabatic, in a steady updraught of strength U cm sec^{-1}, their rates of growth are governed by Eqs. (8) and (15) viz :

$$\frac{d\sigma}{dt}=AU-4\pi B\,n\,\rho\,r^2\frac{dr}{dt} \quad . \quad . \quad . \quad . \quad (8a)$$

and

$$r\frac{dr}{dt}=\frac{1}{C}\left[\sigma-\frac{P}{r}+\frac{Q}{r^3}\right], \quad . \quad . \quad . \quad . \quad (15a)$$

where

$$C=\left[\frac{L^2 M\rho}{K\mathbf{R}T^2}+\frac{\rho\,\mathbf{R}\,T}{DM\,e_s(T)}\right]=\left[9{\cdot}05\,\frac{L^2}{KT^2}+4{\cdot}62\times10^3\,\frac{T}{D\,e_s(T)}\right],$$

with K measured in cal cm^{-1} sec^{-1} °K^{-1}, L in cal g^{-1}, D in cm^2 sec^{-1}, and $e_s(T)$ in mb. The parameters A, B, C are functions of temperature and pressure only.

Eqs. (8a) and (15a), written in terms of finite differences :

$$\Delta\sigma=AU\,\Delta t-4\pi B\,n'\frac{\rho}{\rho_a}r^2\,\Delta r \quad . \quad . \quad . \quad . \quad (8b)$$

and

$$\Delta r=\frac{1}{Cr}\left[\sigma-\frac{P}{r}+\frac{Q}{r^3}\right]\Delta t, \quad . \quad . \quad . \quad . \quad (15b)$$

where n' is now the number of droplets per cm^3, and ρ_a the air density, can be used to calculate σ and r as functions of time (or height above cloud base) in terms of U, n', the nuclear mass m, the temperature and pressure. The results of such a numerical computation are shown in Fig. 1 in which a homogeneous population of sodium-chloride nuclei of $m=10^{-15}$ g is assumed to enter cloud base at 900 mb, 10°C in a concentration of 300 cm^{-3} and to be lifted thereafter in a steady updraught of 1 m sec^{-1}. The calculation shows that the supersaturation attains its peak value of 0·4 per cent after only 10 sec, or 10 m above cloud base; at this stage the droplet radius is 2 μ. Thereafter the supersaturation falls as water vapour is being condensed on the growing droplets more rapidly than it is being released by the cooling air. After 700 sec, when the droplet radius has attained 10 μ, the supersaturation has fallen to 0·07 per cent.

It was found that, after about 200 sec, when the Q/r^3 term in Eq. (15b) becomes negligible, the following solutions of (8b) and (15b)

$$\sigma=\sigma_{200}+\left[P+\frac{ACU\rho_a}{4\pi B\,n'}\right]\left[\frac{1}{r}-\frac{1}{r_{200}}\right], \quad . \quad . \quad (16)$$

$$t=\frac{4\pi B\,n'}{3AU\rho_a}r^3+\left[\frac{P}{AU}+\frac{C\rho_a}{4\pi B\,n'}\right]\frac{1}{r} \quad . \quad . \quad . \quad (17)$$

represent the time variation of σ and r quite accurately.

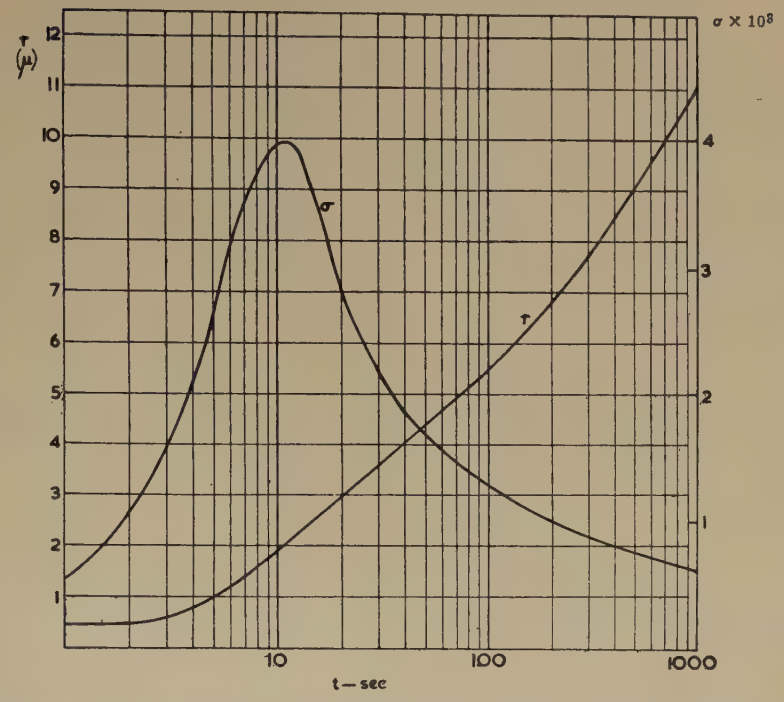

Figure 1. The trend of the supersaturation and droplet radius as functions of time.

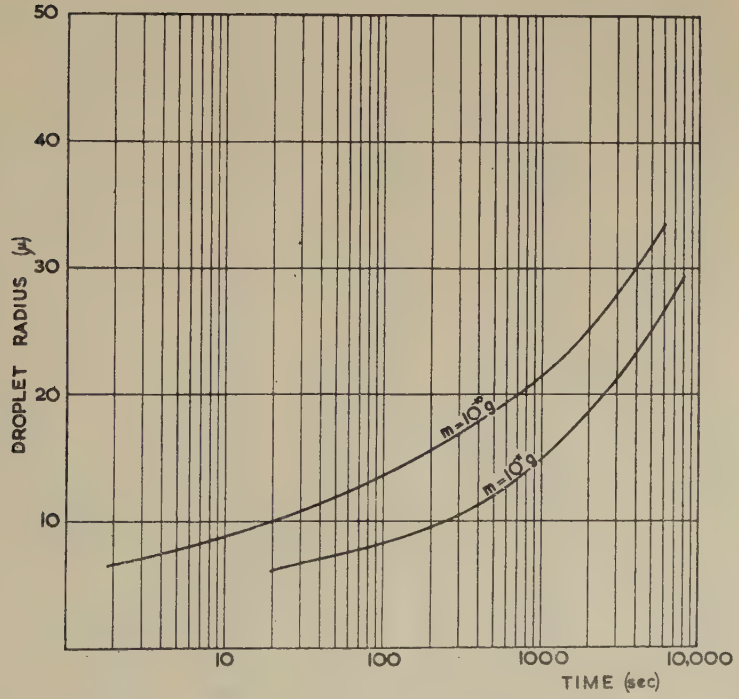

Figure 2. The growth of droplets on giant nuclei containing 10^{-11} and 10^{-10} g of sodium chloride.

4. Growth of giant nuclei in a cloud

We now consider the rate of growth of rare giant nuclei carried up among a population of much smaller nuclei from cloud base, growth in the early stages being assumed to occur by both condensation and coalescence in the core of a rising bubble which does not become diluted by mixing with the environment. We have simplified the problem to the extent of assuming that the much more numerous smaller nuclei are all of the same size so that we can use the results of the computation described in the last section. This assumption is unlikely to affect materially our conclusions, since Howell's calculations indicate that the computed values of supersaturation are not strongly dependent upon the form of nucleus-size distribution.

Our giant nuclei are therefore assumed to enter the cloud base as droplets of saturated salt solution and thereafter to be subjected to the variation of supersaturation with height (or time) shown in Fig. 1. The growth, by condensation alone, of nuclei of mass 10^{-11} g and 10^{-10} g, calculated from the equations of Keith and Arons (1954), is represented in Fig. 2. The total growth rate of a droplet by both condensation and coalescence is given by

$$\frac{dR}{dt} = \dot{R}_c + \frac{Ew}{4\rho}(V - v), \quad . \quad . \quad . \quad . \quad (18)$$

where $\dot{R}_c$ represents the growth rate by condensation, and the second term, the contribution made by coalescence. E is the collection efficiency of the droplet of fall speed V for the smaller cloud droplets of fall speed v, w is the liquid-water content of the cloud, and ρ the density of water. Values of $\dot{R}_c$ for various values of t are evaluated from Fig. 2, and those of $w = \frac{4}{3}\pi r^3 n' \rho$ and v from the radius-time curve of Fig. 1. Being concerned with droplets of $R = 20\text{-}30\ \mu$ colliding with smaller droplets of about half the size, the only available data for E are those of Pearcey and Hill (1957). Evaluating the right-hand side of (18) for various values of t, the increment ΔR in the radius of the drop during the interval Δt was calculated from $\Delta R = \int_t^{t+\Delta t} \frac{dR}{dt}\, dt$. The procedure was repeated step by step for a total period of 1,500 sec during which the particles were lifted 1,500 m above cloud base, and nuclei of mass 10^{-11} and 10^{-10} g attained radii of 17 μ and 28 μ respectively.

At this stage, when the cloud has attained a depth of 1,500 m, the updraught is assumed to be arrested; new bubbles may sustain the cloud but there is no further vertical development. It is now assumed that some of the droplets which have grown upon the giant nuclei while being carried up in the bubbles later become mixed downwards to lower levels in the cloud, growing by coalescence on the way. This may explain the presence of the larger droplets which are found in appreciable concentrations only a few hundred metres above the bases of shallow clouds. The air which carries down the droplets will be subjected to dynamical warming and will continue to move downwards only if it is chilled by partial evaporation of the liquid water which it contains in the form of cloud droplets. The larger droplets with which we are concerned, being much larger than their equilibrium radius appropriate to 100 per cent relative humidity, are unlikely to evaporate in the cloudy air. Evaporation of the water in the downward-moving eddies, although partly offset by mixing with the surrounding cloudy air, will cause the water content of the bulk of the cloud to become considerably less than that calculated on the basis of adiabatic ascent from cloud base. Measurements made from aircraft indicate values of a few tenths of gramme per m^3.

Assuming a mean water content of 0·5 g/m^3 composed of droplets of mean-volume radius 7 μ, our calculations show that droplets originating on nuclei of mass 10^{-10} g would grow from 28 to 50 μ radius in 35 min, making a total time of one hour. In the same time

and under the same conditions, nuclei of 10^{-11} g would give rise to droplets of radius 30 μ. In computing this later growth of the larger droplets by coalescence with much smaller ones, we have used Langmuir's values for collection efficiency suitably modified to allow for the finite size and fall speed of the smaller droplet.

The calculations just described were made for an updraught of 1 m/sec corresponding to a rather feeble cloud. If the updraught speed were doubled, the supersaturation and droplet growth would increase more rapidly with time, with the result that the larger droplets growing in the ascending bubble would require about one-third less time to reach the same size, but the depth of the cloud would be increased by about one-third.

5. Discussion

These calculations show that a cloud only about $1\frac{1}{2}$ km deep, providing it lasts for about one hour, may produce droplets of radii 30-50 μ if it is provided with hygroscopic nuclei of mass 10^{-11} to 10^{-10} g. In a recent paper East (1957), assuming that droplets of radius 18 μ exist a few hundred metres above cloud base, shows that on being lifted along a saturated adiabatic, they may grow to 50 μ by condensation and coalescence after a further 20-30 min, and suggests that in consequence it may be unnecessary to postulate the existence of giant nuclei to initiate a coalescence shower. However, when one enquires into the origin of the 18 μ droplets from which East starts his calculations, it appears (see Fig. 2) that their formation within a few hundred seconds can only be explained in terms of growth upon nuclei of mass about 10^{-10} g.

Over the oceans, in areas of spray formation, Woodcock (1953), and Moore and Mason (1954) have found nuclei of mass 10^{-10} g in concentrations of order 10/l, those of 10^{-11} g being about ten times more abundant. Such concentrations of giant salt nuclei are only just sufficient to explain the observed concentrations of large droplets in some small cumuli bearing in mind that the spatial concentrations of nuclei carried up in ascending bubbles will become diluted by subsequent mixing throughout the bulk of the cloud. Over southern England and other areas close to the sea, the numbers of giant hygroscopic nuclei may sometimes be comparable to those occurring over the oceans, but it would seem highly desirable to measure *simultaneously* from aircraft the concentrations of both the large cloud droplets and the giant nuclei, in order to determine whether the latter may account for the former.

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The use of balloons in atmospheric pollution research

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SUMMARY

An experimental technique has been established for measuring the rise of a plume from a stack, the ground-level concentrations produced and the rates of deposition of dust emitted in the plume. It can be applied over distances where plumes from power stations are not normally visible. It can be used to give a result before weather and station conditions have changed appreciably and therefore produces information which can conveniently be compared with theoretical estimates.

1. INTRODUCTION

The expenditure of money on the prevention of atmospheric pollution by dust and gas emitted from power-station chimneys is so great that it is essential to have a firm scientific knowledge of the behaviour of dust and gas after emission from the chimneys so that the expenditure may be made to the best purpose. There are several theoretical methods which enable forecasts to be made of the behaviour of effluents under various conditions, but these are to some extent in conflict and it is generally agreed that they require practical confirmation before they can be used with confidence.

One of the most uncertain factors in the theory of atmospheric pollution is the height to which the plume rises above the chimney top owing to the buoyancy of the hot gas. Another factor which has not received adequate theoretical treatment is the rate at which dust separates from the plume during its period of rise due to buoyancy. The first factor is of major importance because it has a powerful effect on the concentrations of dust and gas subsequently produced at ground level. The second factor is of importance because the dust which separates quickly from the plume does not benefit from the plume rise due to buoyancy.

Direct measurements of dust deposition have been made in an attempt to achieve an empirical result which could be compared with the theoretical predictions. These measurements are of limited value because the rate of deposition cannot be measured in a short enough time to obtain a result while one set of conditions persists, and are spoiled by the inclusion of dust from other sources than the source under study. It is true that a particular dust can be labelled by tracers but the techniques are very complicated to apply.

It has therefore been decided to develop a technique which will enable estimations of dust deposition and gas concentration to be made in such a way that the results can be correlated with the important variables. These are the height of the chimney, the quantity and quality of the dust and gas emitted, the wind speed, the degree of turbulence in the atmosphere, the total amount of heat emitted, the emission velocity and other subsidiary factors such as the nature of the surrounding countryside. It is the object of this paper to describe the development of the technique and to give some results to illustrate it. It is hoped to present informative results in a later paper.

2. THE NATURE OF THE TECHNIQUE

The technique is based upon the use of a balloon to represent either a particle of dust or a quantity of gas. The balloon is filled with a gas, such as hydrogen or methane, and its weight is adjusted until either its free-falling speed is zero or has some fixed value.

n the first case it will represent a quantity of gas; in the second case it will represent dust particle of the specified free-falling speed.

Initial tests were made with small meteorological balloons with a circumference of 6 or 48 in. They were inserted into the base of the chimney of Little Barford Power tation and were carried out of the chimney by the flue gases. It was found that these alloons could successfully traverse the chimney on certain occasions, but it was also ound they had certain serious drawbacks. As a result of these tests the following require- nents for a suitable balloon were established :

(1). The balloon must have good resistance to the heat of the stack. The stack emperature varies in different cases from 250°F to 350°F.

(2). The balloon must have good resistance to the diffusion of gas through its nvelope. If gas escapes from the balloon the buoyancy changes and this again upsets nore or less rapidly the initial adjustment of free-falling speed.

(3). It is desirable that the balloon should be unaffected by any radiation falling on it. f the balloon absorbs radiation its temperature increases, and this has a powerful effect n modifying the initial adjustment of free-falling speed.

(*a*) *Development of a suitable balloon*

It was soon found that natural rubber balloons were not suitable. At higher stack emperatures they invariably burst, and at the lower temperatures, although a fair propor- ion came out of the chimney, the rubber had been so affected by the heat that the balloon vas slightly larger when it came out, which meant that the original adjustment of free- alling speed had been invalidated.

Light gases diffuse to a varying degree through thin membranes, and hydrogen liffuses three or four times more quickly than methane. Fig. 1 shows the change in uoyancy of rubber balloons of different sizes and make containing both hydrogen and nethane. Because of the more rapid loss of buoyancy, a hydrogen-filled rubber balloon ould not be used to simulate gas or particles having a low free-falling speed.

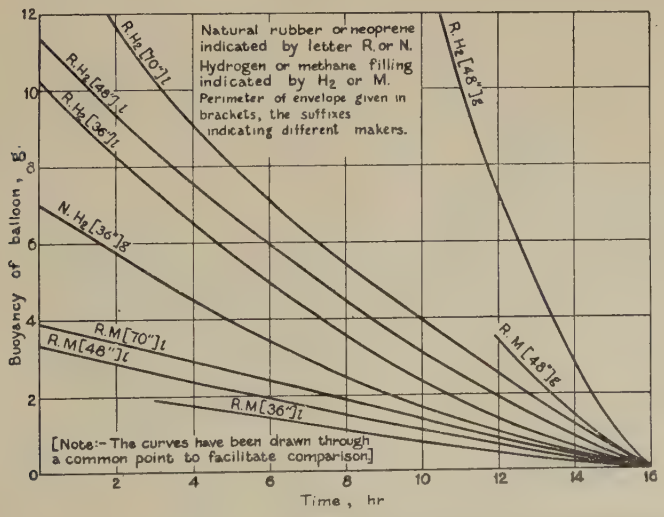

Figure 1. Variation in buoyancy of natural rubber and neoprene gas-filled balloons.

It was found that the resistance to diffusion of rubber balloons could be considerably improved by using a double balloon. A balloon, enclosed in a second balloon, was filled with hydrogen or methane and the interspace was filled with air. The rate of loss of buoyancy initially was zero, the gas escaping from the inner balloon being held by the outer balloon, but increased as an appreciable concentration of hydrogen or methane was built up in the interspace (Fig. 2). It can be seen therefore that the performance of a double balloon was considerably better than that of a single balloon with respect to changes in buoyancy for a period of two or three hours. Although this overcame one difficulty, making a balloon suitable for work at atmospheric temperatures, it did not overcome the difficulty of low resistance to heat.

Rubber balloons absorb radiation and cannot be used when the sun is shining.

Many materials were considered for the fabrication of a suitable balloon, including neoprene rubber, cellophane, polythene and terylene. Terylene film had the best available resistance to diffusion, and it had excellent heat resistance. It had the disadvantages that

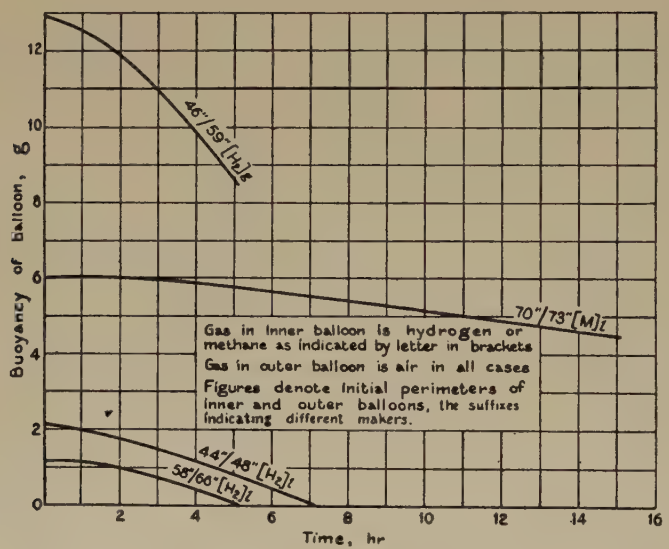
Buoyancy of balloon, g
Time, hr
46"/59"[H2]5
70"/73"[M]1
44"/48"[H2]1
58"/66"[H2]1
Gas in inner balloon is hydrogen or methane as indicated by letter in brackets
Gas in outer balloon is air in all cases
Figures denote initial perimeters of inner and outer balloons, the suffixes indicating different makers.

Figure 2. Variation in buoyancy of double rubber gas-filled balloons.

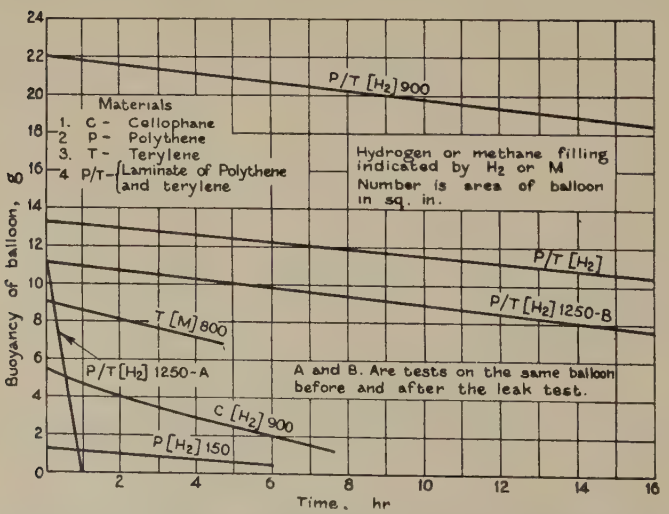
Buoyancy of balloon, g
Time, hr
P/T [H2] 900
Materials
1. C - Cellophane
2 P - Polythene
3. T - Terylene
4 P/T - Laminate of Polythene and terylene
Hydrogen or methane filling indicated by H2 or M
Number is area of balloon in sq. in.
P/T [H2]
P/T [H2] 1250-B
T [M] 800
P/T [H2] 1250-A
A and B. Are tests on the same balloon before and after the leak test.
C [H2] 900
P [H2] 150

Figure 3. Variation in buoyancy of gas-filled balloons made of various types of plastic sheet.

it was liable to have small pin-holes and was extremely difficult to fabricate. On the other hand polythene had a relatively poor resistance to diffusion but it was extremely easy to fabricate.

Finally, a laminate of terylene and polythene was obtained which combined the advantages of both materials. The terylene gave good diffusion resistance and the polythene enabled fabrication to be carried out by heat sealing. It also sealed up any pin-holes in the terylene. Since this material is inelastic it must be made in the form of a bag and must be only partially filled with hydrogen, so that there is space left for expansion. This was in itself an advantage. If the balloon fabric were in tension any change in the pressure would upset the initial adjustment, but since the balloon is invariably slack, the gas is always at atmospheric pressure and this source of error is eliminated. The plastic material was rather heavier than the normal rubber balloon and so the balloon had to be larger. This offsets to a large extent the improved resistance to diffusion because of the larger surface area. The diffusion rates of various plastic balloons are shown in Fig. 3.

The plastic balloons successfully met two of the requirements given above. Although the material is transparent to visible light it did not comply with the third requirement as it had appreciable infra-red absorption. Completely reliable results cannot therefore be obtained in sunny conditions, until some experimental work has been carried out to determine the correction to be made to the weight of the balloon. However, when the sky is overcast the error due to radiation effect is negligible and balloons can be successfully used under these conditions.

After the balloon had been partially filled with hydrogen it was attached to one arm of a balance and enclosed in a draught-free box. The buoyancy of the balloon was adjusted by adding string until the weight corresponding to the required free-falling speed was obtained.

If the balloon was not completely gas-tight, balancing was impossible. Balloons rejected on these grounds were later subjected to a leak-finding test (see Fig. 3) by filling the balloon with air bubbled through ammonia and then placing the balloon in a transparent chamber containing an atmosphere of hydrochloric acid gas. Leaks were quickly made evident by the formation of ammonium chloride.

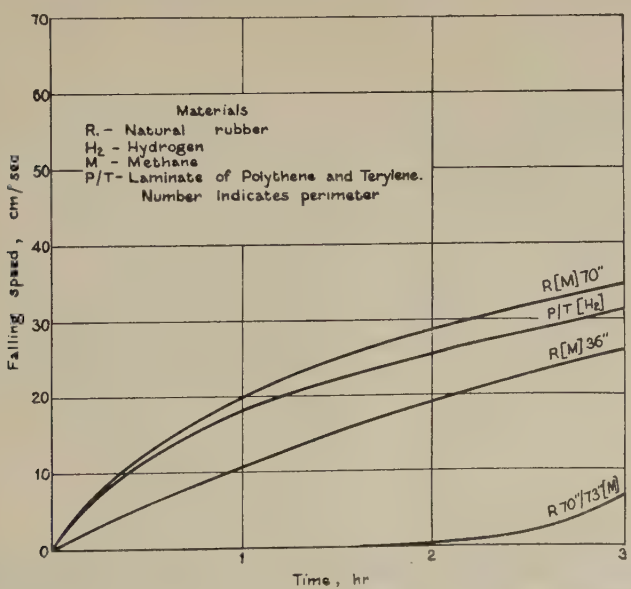

Figure 4. Variation of falling speed with time for different types of balloons, initially just buoyant.

The balloon was then introduced either into the chimney or into the duct leading to the chimney, according to circumstances. The buoyancy of the balloon varies with time owing to loss of gas so that a balloon which is just buoyant initially descends slowly after a while. Theoretically, $v \propto w^{\frac{1}{3}}$ where v is the falling speed and w is the increased weight or loss of buoyancy. Free-falling speeds of polythene/terylene balloons and rubber balloons for different weights were determined by experiment, and using the known diffusion rates the variation of falling speed with time is shown in Fig. 4. The greater resistance to diffusion of the polythene/terylene balloon over the rubber balloon is also shown in Fig. 4 when it is remembered that hydrogen diffuses much more quickly than methane.

(*b*) *Development of a technique for following the path of a balloon*

After the balloon had been emitted from the chimney it was necessary to follow its flight so that quantitative information on its behaviour could be recorded. It was desirable to note the path of the balloon through the air in addition to the point at which it reached the ground, since the actual path of a balloon with zero buoyancy would indicate the plume rise due to buoyancy and if this were known it would greatly facilitate the theoretical analysis. The use of theodolites, cameras and radar was considered in the initial stages. A test was carried out with a 3-cm radar set, which was kindly loaned and operated by Messrs. Decca Radar Ltd. A radar target made of aluminised paper was hung on each balloon and it was found that single balloons could be followed by radar up to a distance of about five miles. The angular height of the balloon could also be approximately indicated by the radar equipment. The use of radar was initially considered because it might make it possible to follow a large number of balloons at one time. However, the idea was abandoned because it was felt that the extra cost and elaboration of the radar technique to give accurate height measurement were not justified, and since it would not have been at all easy to follow a number of balloons.

The technique ultimately used was to fix a naval rangefinder on a mounting which enabled the rangefinder to be moved in elevation and azimuth. Since this apparatus gave range, elevation and azimuth, it gave a fix for the balloon at any given time. In order to facilitate the taking of the successive readings a potentiometer was connected to the mounting so that a change of elevation produced a changing D.C. voltage and this was recorded on a pen-recorder. This automatically gave a plot of elevation against time. It was possible to record range in addition, by specifying a number of fixed ranges, for example, 400 yd, 600 yd, 800 yd, etc. As the balloon reached each of these distances a switch was operated, short-circuited the potentiometer signal and so gave a clear indication on the recorder chart. Readings of azimuth which were of minor importance were written down separately and collated afterwards. This unit was mounted on the roof of the power station and so had a very clear view under most circumstances. The method has the advantage that, not only does it plot the path of the balloon, but that it gives a very reliable measure of wind speed (a cup anemometer a few feet above the roof of the power station was found to under-read by 60 per cent). This reliable and detailed information about mean wind speed and its variation from moment to moment has proved of great value in assessing the results obtained.

(*c*) *Validity of the method*

(*i*) *Theoretical considerations.* It may be argued that the balloon is so different in physical size from either the dust particle or the gas molecule that it cannot give a true indication of the behaviour of either. However, the comparison to make is between the

size of the phenomenon and the size of the balloon, rather than between the balloon and the dust particle. The atmospheric eddies which affect the motion of the balloon and the dust particle cover a range of sizes from the mean free path of the gas molecules up to eddies with diameters of hundreds of yards. It is true that the balloon cannot respond to the small-scale eddies, but it is certain that the effects produced by the larger eddies are very much more marked than the effects produced by the smaller ones, except very close to the ground, so that while the motion of the balloon is different from the motion of the dust particle in its fine detail, it is very closely similar in its main outline. It is therefore thought that while it is true in principle that the behaviour of the balloon and the dust particle are not identical, in practice the error involved is extremely small compared with the uncertainties both of theory and of measurement.

Somewhat different considerations apply as soon as the balloon is in the neighbourhood of the ground. When the balloon is simulating dust the parallel between the two cases remains close in that both entities are normally trapped by contact with the ground. When the balloon is simulating gas an error is involved because the balloon is likely to be trapped by trees and fences on the earth surface but a gas particle is not. This means that the results obtained with balloons in the study of gas must be treated with caution after the gas has achieved contact with the ground. However, at the present stage of the investigation, the movement of the gas before it reaches the ground is of considerably more interest and importance than its subsequent motion.

(*ii*) *Comparison with smoke puffs.* There are certain practical difficulties in the use of balloons. If a balloon is balanced to be exactly buoyant at one temperature it will be exactly buoyant at any other temperature, provided the balloon and its surrounding gas have both reached the same temperature. However, while the balloon is at a different temperature from its surrounding gas, an error exists. As the balloon is passing up the chimney it will be cooler than its surrounding gas. It will therefore be emitted with a certain free-falling speed. However, since at the moment of emission the plume is being bent sideways by the effect of the wind, the extra density of the balloon will cause it to be flung upwards. These two factors tend to cancel out. The gas temperature of the plume drops rapidly immediately it has left the chimney and, within a few seconds, balloon temperature and gas temperature are the same. Subsequently the gas probably cools slightly more quickly than the balloon, and the balloon will tend to have a slight upward motion until it achieves ambient temperature. This effect must necessarily be small compared with the overall buoyancy of the plume, and while it may cause the balloon to be slightly higher in the plume than it otherwise would be, it cannot have a major effect on the total plume rise due to buoyancy as indicated by the balloon.

In order to check as far as possible that the three sources of error mentioned do not have a serious net result, the movement of balloons has been compared with the movement of smoke. Since the plume normally emitted from a modern power station is not clearly visible, smoke was deliberately introduced into the chimney by means of an oil-smoke generator very kindly loaned by the Director of the Chemical Defence Experimental Establishment, Porton. Puffs of smoke were emitted from the chimney at the same time as balloons and during the period that the smoke remained visible the balloon was observed to stay with the plume. The distance concerned was about $\frac{1}{4}$ mi, and it is considered that the transient errors mentioned above would have had their full effect in this distance.

3. The type of result obtained

The results which follow are insufficient in number to be discussed and are given merely to illustrate the technique.

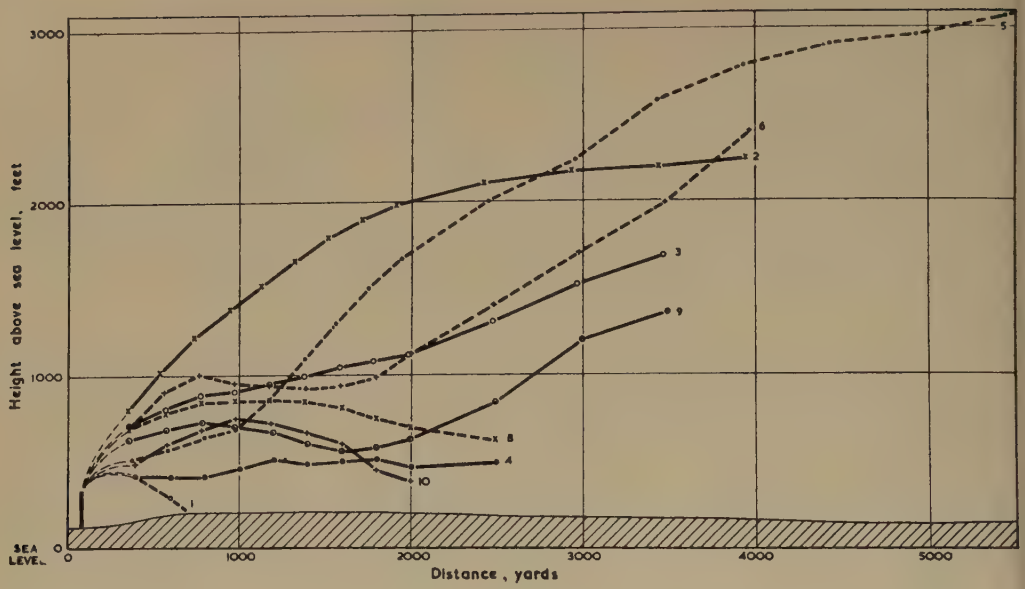

Figure 5. Vertical section of balloon tracks, 8 August 1956, 1200-1600 hr. Weather, sunny. Mean wind, 22·5 ft/sec.

Figure 6. Vertical section of balloon tracks, 16 August 1956, 1300-1500 hr. Weather, cloudy. Mean wind, 22·5 ft/sec.

(*a*) *Observation of plume rise due to emission velocity and buoyancy*

Figs. 5, 6, 7 and 8 show the results obtained with balloons used to simulate quantities of gas. The tests were carried out at Earley Power Station. The balloons were emitted over a period of several hours and so indicate the time mean profile of the plume and not its instantaneous behaviour. It is possible, by taking the average heights of all the balloons at different distances, to estimate the total plume rise, viz., 500 ft on 8 August and 600 ft on 16 August. Knowing the plume rise, the maximum ground-level concentration can be estimated using Sutton's diffusion formula, assuming that the wind direction was constant. By taking the heights of all the balloons at any particular distance a distribution pattern can be plotted against height. Taking the azimuth distribution of all the balloons it is

Figure 7. Plan of tracks taken by balloons, 8 August 1956, 1200-1600 hr.

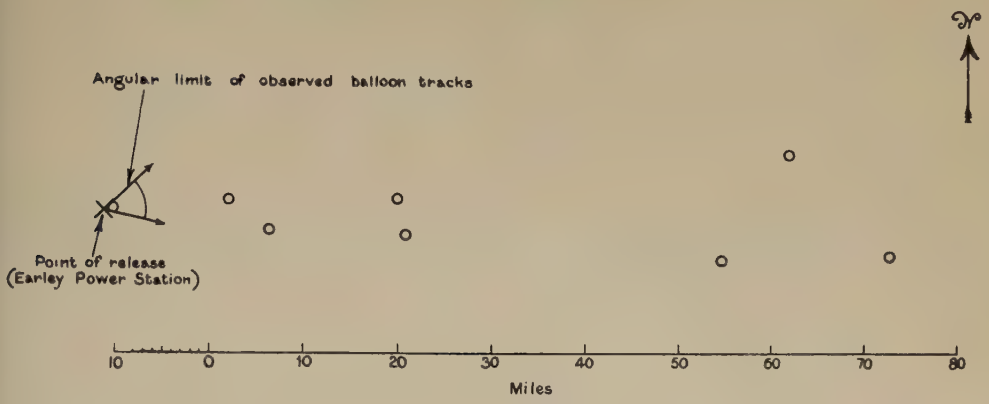

Figure 8. Recovery points of balloons released on 8 August 1956.

possible to plot a distribution pattern at right angles to the plume. This makes the same assumption as Sutton's work, that horizontal and vertical distributions are independent. From these results a mean ground-level concentration can be calculated at any particular distance from the power station and this will take account of the variations in wind direction which occurred during the test run. It is of interest that at the distance of maximum concentration, the concentration averaged over the test period is a quarter of the maximum concentration.

(*b*) *Application to dust dispersion*

The second result, which is given in Fig. 9, represents the other end of the scale. It was obtained at Little Barford Power Station. It shows the deposition pattern obtained with balloons adjusted to represent very coarse dust with a free-falling speed of 170 cm/sec. Again, in order to make this plot it is necessary to assume that variations in azimuth do not affect the behaviour of the dust particle. It can be calculated that for very heavy dust of this kind the effective height of the chimney is very little greater than the chimney height itself although the plume rise must have been considerable. This illustrates the fact that heavy dust leaves the plume relatively early and does not benefit from the plume rise due to buoyancy.

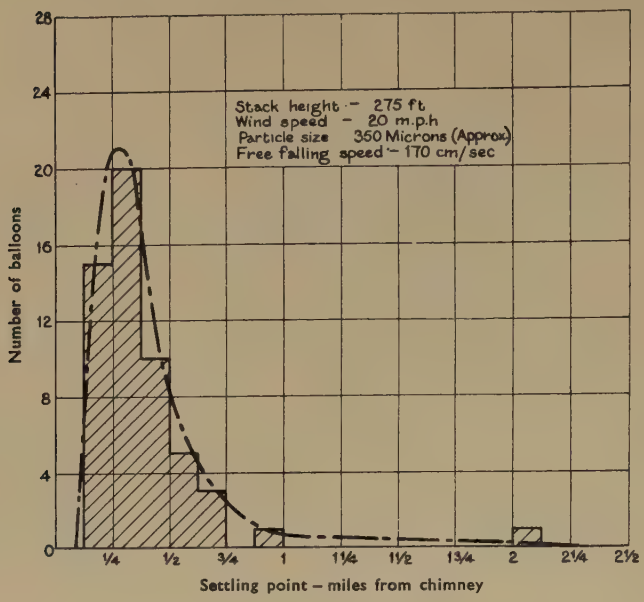

Figure 9. Dust deposition test results.

(c) *Further application of the balloon technique*

Using balloons of zero buoyancy, the technique has also been applied to study wind flow, e.g., over buildings and valleys, under cloudy conditions. In this case, being a low-temperature application, rubber balloons have proved to be quite satisfactory.

Acknowledgments

This paper is published by permission of the Director of the Research Laboratories, Central Electricity Authority.

We are grateful to the I.C.I. for the supply of polythene/terylene laminate; to the Paper Goods Manufacturing Co. Ltd., Sutton, for their assistance in the development and production of the plastic balloons; to the staffs of Little Barford and Earley Power Stations for their co-operation; and to our colleague Mr. B. L. Davies for his assistance with the experimental work.

551.508.25 : 551.521.61

A simple grating spectrometer for sky-emission studies

By R. M. GOODY
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(Manuscript received 23 July 1957)

1. Introduction

Low-temperature emission spectrometry is inevitably energy limited and, for a given detector, the minimum slit width for a given signal-to-noise ratio varies with the factor

$$\left(\frac{T\frac{d\theta}{d\lambda}A}{R}\right)^{-\frac{1}{2}}$$

where T is the transmission of all optical components,

$\frac{d\theta}{d\lambda}$ is the dispersive power of prism or grating,

A is the height of the prism or grating,

R is the focal length to diameter ratio of the detector collecting mirror.

An instrument has been constructed at the Imperial College Field Station, Silwood Park, Ascot which gives the maximum resolution compatible with reasonably low cost.

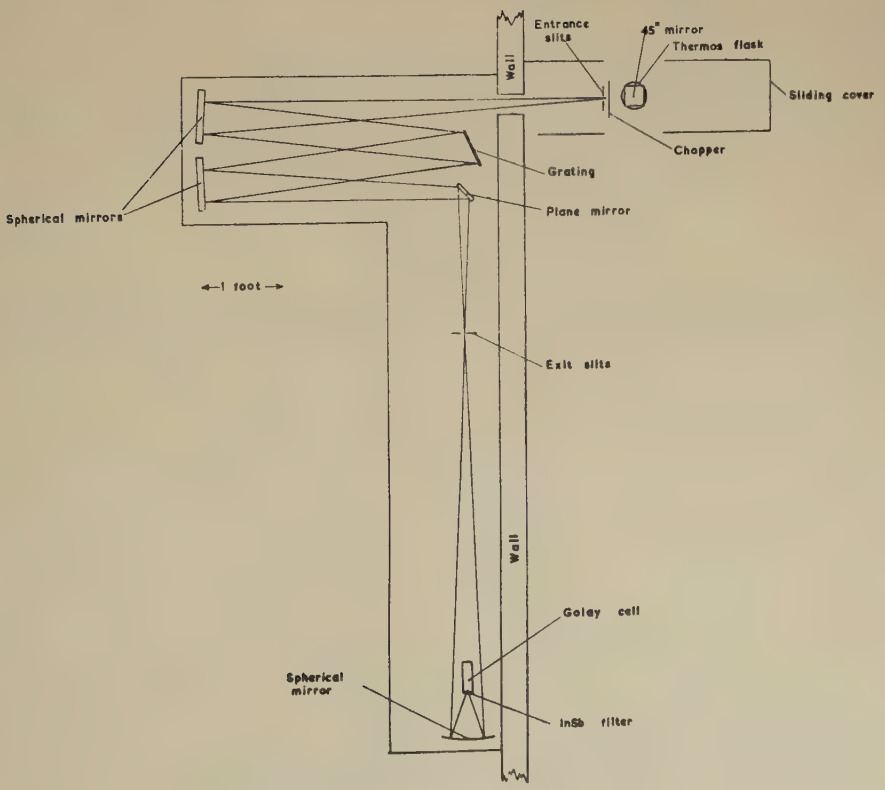

Figure 1. General layout of the spectrometer.

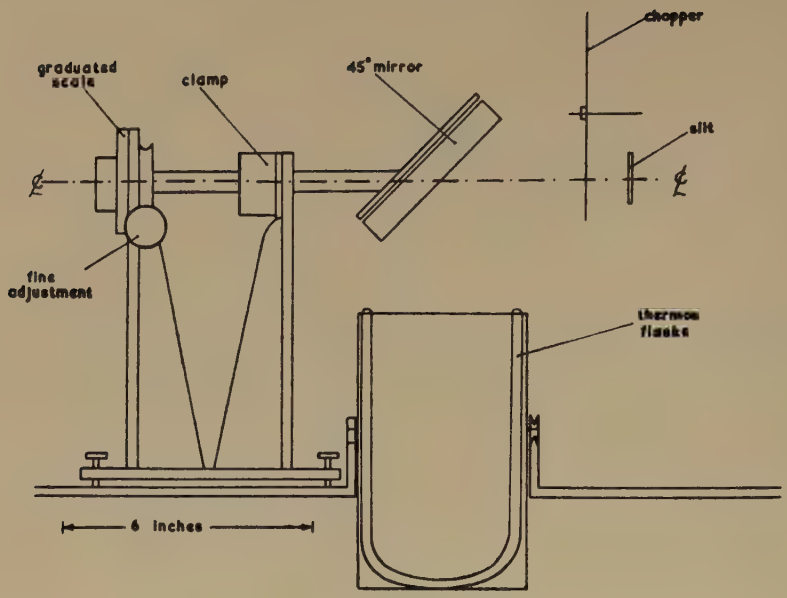

Fig 2. External optics.

Large dispersion is best obtained with a grating. For work of this type where high resolution is impossible, there is no obvious limitation to the size of grating which can be used. Dr. Sayce, of the National Physical Laboratory, offered to phase four 6 in. × 6 in. Merton gratings, but consideration of the cost of the other optical components led to the decision to try to get the best out of one such grating.

The Merton grating is of high optical quality. The grating used has 2,500 lines/in. and a blaze angle of 60° at 1,000 cm^{-1}; 60 per cent of radiation of this frequency goes into the first order. We were able to establish when calibrating with a Nernst filament that the grating was capable of resolving at least 0·25 cm^{-1} even without careful adjustment; Professor Price (private communication) informs us that in fact 0·1 cm^{-1} can be resolved with a similar grating.

2. Description

A modified Ebert system has been employed (Fig. 1) with grating and slits at different distances from the spherical mirrors, so that the entrance slit can be placed outside the observing hut under a movable housing to avoid auxiliary focusing outside the spectrometer.

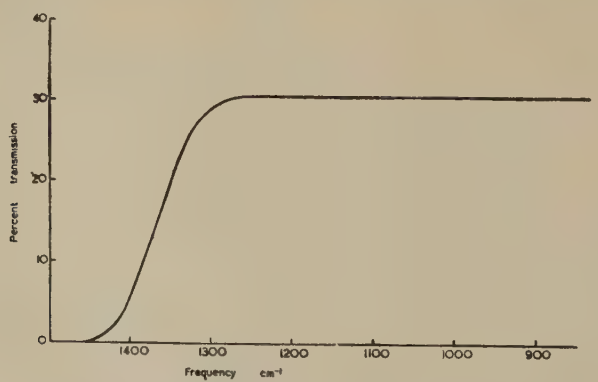

Figure 3. Transmission of a 50 μ indium-antimonide filter.

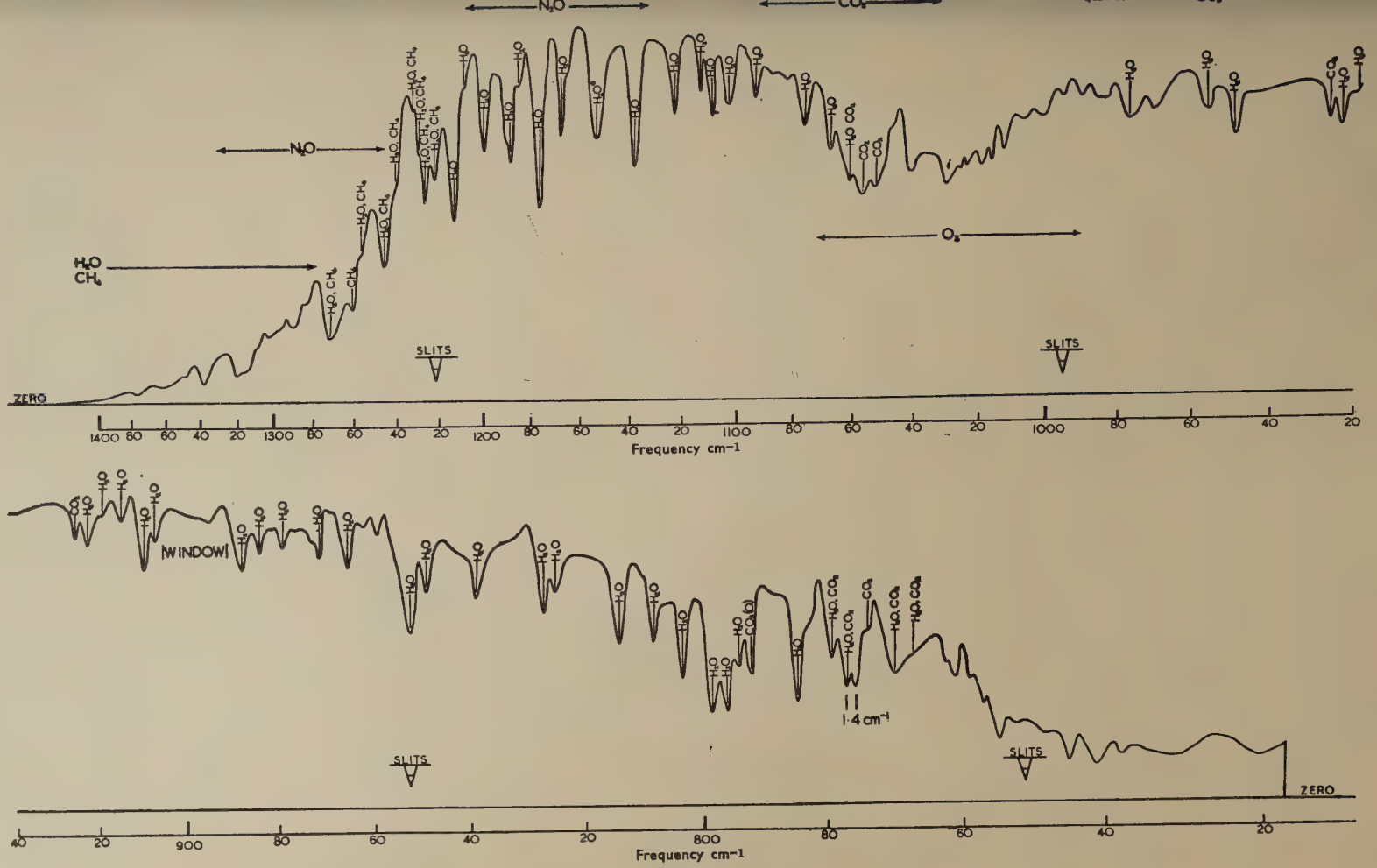

Figure 4. Sky emission-spectrum for three air masses 19 April 1957 at 1400 hr.

Following the entrance slit are a shutter operated by a Bowden cable, a polished aluminium chopper and a plane mirror mounted at 45° to a horizontal axis (Fig. 2). By rotating about this axis the spectrometer can view the sky at any desired air mass or can look downwards into one of two Dewar flasks filled with water and liquid air respectively and mounted on a horizontal slide.

The detector is a Golay cell and the collecting mirror has an aperture $f/1$. A larger aperture would serve no purpose on account of the high reflectivity for oblique rays of the indium antimonide filter used to separate off higher orders of diffraction.

The transmission curve of the indium-antimonide filter is shown in Fig. 3. In this frequency range no radiation of higher orders could be detected with a Nernst filament as a source, and since sky radiation has a far greater proportion of long-wave radiation the filter is completely effective for frequencies from 1,400 to 700 cm^{-1}.

The filter was 50 μ thick and was made for us by Dr. G. A. Avery of the Radar Research Establishment. The refractive index of indium antimonide is about 3·75 (Roberts and Quarrington 1955) so that almost all of the loss beyond 1,300 cm^{-1} can be attributed to reflection and could therefore be avoided by blooming.

3. Performance

Fig. 4 shows a tracing of an emission spectrum from 1,400 to 700 cm^{-1} recorded on 19 April 1957 at 1,400 hr with three air masses. The slits were 3·17 mm wide, corresponding to 3·7 cm^{-1} at 1,325 cm^{-1} and to 0·98 cm^{-1} at 741 cm^{-1}. At about 780 cm^{-1} a resolved doublet of spacing 1·4 cm^{-1} can be seen.

The lines were identified from the solar spectra of Shaw, Oxholm and Claassen (1952). Individual lines and blends of two or three lines of water vapour, methane and carbon dioxide can be identified. Unresolved bands of carbon dioxide and nitrous oxide are also marked. The 1045 cm^{-1} ozone band shows as a major feature of the spectrum and some of the structure begins to be resolved.

At a certain frequency ν let

C_ν = energy emitted by the chopper blade.

S_ν = energy emitted by the sky,

W_ν = energy emitted by the water surface,

P_ν = energy emitted by entire optical path including the 45° mirror to entrance slit,

τ_ν = transmission of entire optical path including the 45° mirror to entrance slit.

The deflection of the recorder is proportional to the difference of energy when the chopper blade is covering and not covering the slit. Let D_s, D_0 and D_w be the deflections recorded when the instrument looks at the sky, the liquid air and the water surface. Since the liquid-air emission is effectively zero we have :

$$S_\nu \tau_\nu + P_\nu - C_\nu = \alpha D_s$$

$$P_\nu - C_\nu = \alpha D_0$$

$$W_\nu \tau_\nu + P_\nu - C_\nu = \alpha D_w$$

where α is a constant. Hence,

$$\frac{S_\nu}{W_\nu} = \frac{D_s - D_0}{D_w - D_0}$$

Thus by taking subsequent spectra of the sky, the water and the liquid air, the sky and water emissions can be compared without corrections. The water is effectively a black-body radiator, and its emission can be computed from its measured temperature. This simple relation is only possible because the angle of reflection and therefore the absorption and emission of the 45° mirror are constant. A slight complication arises because the liquid air shows some reflection, but a small correction can be made for this.

The energy emitted from the chopper blade is a combination of true emission and reflection om the entrance slit jaws and from inside the instrument. This stays sufficiently constant roughout a series of observations. It would be better to use a tilted-mirror chopper which flects radiation from a constant temperature source into the instrument, as is done by Sloan, aw and Williams (1955), but the design of the instrument did not leave sufficient space for is feature to be included.

4. Proposed investigations

W. T. Roach, in this laboratory, has investigated atmospheric absorption and emission ectra in the 1,000 cm^{-1} region at low resolution and has concluded that there is probably a rge background contribution from non-gaseous constituents. With the grating spectrometer it ill be possible to separate line absorption and background absorption with certainty. Reference the spectra of Migeotte, Neven and Swensson (1956) shows that between 906 and 892 cm^{-1}, r example, there are no gaseous absorption lines. This window is marked on Fig. 4 and there , of course, no difficulty in sampling radiation exclusively from this region.

We have also recorded the band area of the 1,045 cm^{-1} ozone band for the purpose of inferring e vertical distribution of this gas (Goody and Roach 1956). With this equipment it will be ossible to observe the change with air mass of the emission in several different parts of the band, hich, having different absorption characteristics, will each give partially independent solutions r the ozone distribution.

5. Improvements in performance

A cone-channel condenser (Williamson 1952) in front of the Golay cell would allow the se of longer slits with proportionately more energy. Owing to the method of construction of the ell, however, this would have to be attached in the course of manufacture. Alternatively, an nage slicer (Benesh and Strong 1951) used at the exit slit should give somewhat similar results.

If the filter were bloomed with a film of germanium the transmission in the centre of the ange of interest could be increased by a factor of two or more. Such a filter is under construction t the time of writing.*

These two improvements together should make possible an increase of two in the resolving ower, giving an average resolution of about 1 cm^{-1} over the range 1,400 to 700 cm^{-1}. This hould permit the resolution of the carbon dioxide bands shown in Fig. 4.

Acknowledgments

I am indebted to Mr. D. A. Rogers for constructing all the equipment and doing much of he detailed design, and to Dr. D. G. Avery for supplying the indium-antimonide filter.

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* *Added in proof*: Dr. R. C. Smith of the Royal Aircraft Establishment has succeeded in blooming a filter for us with ad chloride. The transmission is now on the average more than 60 per cent, and many carbon-dioxide lines have been corded. Near 780 cm^{-1} a doublet with a separation of 1·1 cm^{-1} has been resolved.

551.510.4

The hydrogen content of atmospheric air at ground level

By E. GLUECKAUF and G. P. KITT
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(Manuscript received 4 June 1957)

SUMMARY

A method is described for the separation and determination of hydrogen in air samples of approximately 200 litres N.T.P. by volume. The hydrogen is fractionally distilled from the bulk of the sample, oxidized to water by passing over heated copper oxide and finally condensed in a trap cooled in liquid oxygen. After freeing from the contaminating gases by pumping, the water is released from the trap and measured as vapour in a small calibrated volume.

Although the accuracy of the method in its present form is not better than ± 10 per cent, the results seem to indicate that the hydrogen content of atmospheric air is a varying quantity.

The average hydrogen content of air at ground level is 0·6 parts per million.

1. INTRODUCTION

Very little accurate work has been carried out on measuring the hydrogen content of the atmosphere. Phillips (1895) detected no hydrogen at a height of 7,000 ft above sea level, but H. and E. Erdmann (1911) found it at a height of 800 m. In 1898, Gautier calculated air to contain from 110-180 parts per million but spectroscopic observations by Lord Rayleigh in 1902 led him to an estimate one-sixth of that of Gautier. Leduc's observations on the density of air (1902-1903) also indicate that Gautier's estimate is too high. In 1909 Claude found air to contain less than 10 parts of hydrogen per million.

The International Critical Tables, 1926 and the *Handbook of Physics and Chemistry*, 1956-57 give a figure of 100 parts per million, but a lower figure of 0·5 parts per million is given in the *Smithsonian Physical Tables* 1954 and is quoted by Goody in 1954.

At present it would appear that the most reliable figure available was obtained by the separation and estimation of hydrogen obtained from crude neon manufactured from air and containing all the hydrogen present in air. This work was carried out at the Linde works, Hamburg, and the figure obtained, 0·5 ± 0·05 parts per million, was privately communicated to F. A. Paneth in 1937.

The work described in this paper was started in 1951, making use of a fractionating still used by the authors for determining the krypton and xenon contents of atmospheric air. Although it has not been possible to bring it to a satisfactory conclusion, it is considered that the results obtained to date are worth putting on record.

Essentially the method consists of (i) the liquefaction of about 200 litres N.T.P. of uncontaminated atmospheric air; (ii) the separation of the hydrogen from the higher boiling point gases, including methane of boiling point — 161·5°C, by fractional distillation; (iii) the treatment of the hydrogen-nitrogen mixture passing from the still in a silica tube containing copper oxide heated to 800-900°C to convert the hydrogen to water vapour; (iv) the trapping of the water vapour in a specially modified and calibrated McLeod gauge in which the water can be measured subsequently as water vapour, and (v) the measurement of the amount of gas sample taken.

Although much progress has been made in recent years in the techniques of manipulating small volumes of gas, the quantitative handling of 0·1 to 0·2 ml N.T.P. of hydrogen in the presence of large quantities of nitrogen gives rise to many difficulties. The apparatus used was checked by introducing small quantities of hydrogen with large quantities of nitrogen but, although many of the results from these runs were accurate to ± 10 per cent, occasionally errors of up to ± 20 per cent were obtained.

2. APPARATUS

(*a*) *Fractionating columns*

(i) *Glass apparatus.* A glass fractionating still with a boiler capacity of 250 ml was used for the series of runs made in 1951. This is illustrated in the paper published by the authors in 1956 on the krypton and xenon contents of atmospheric air.

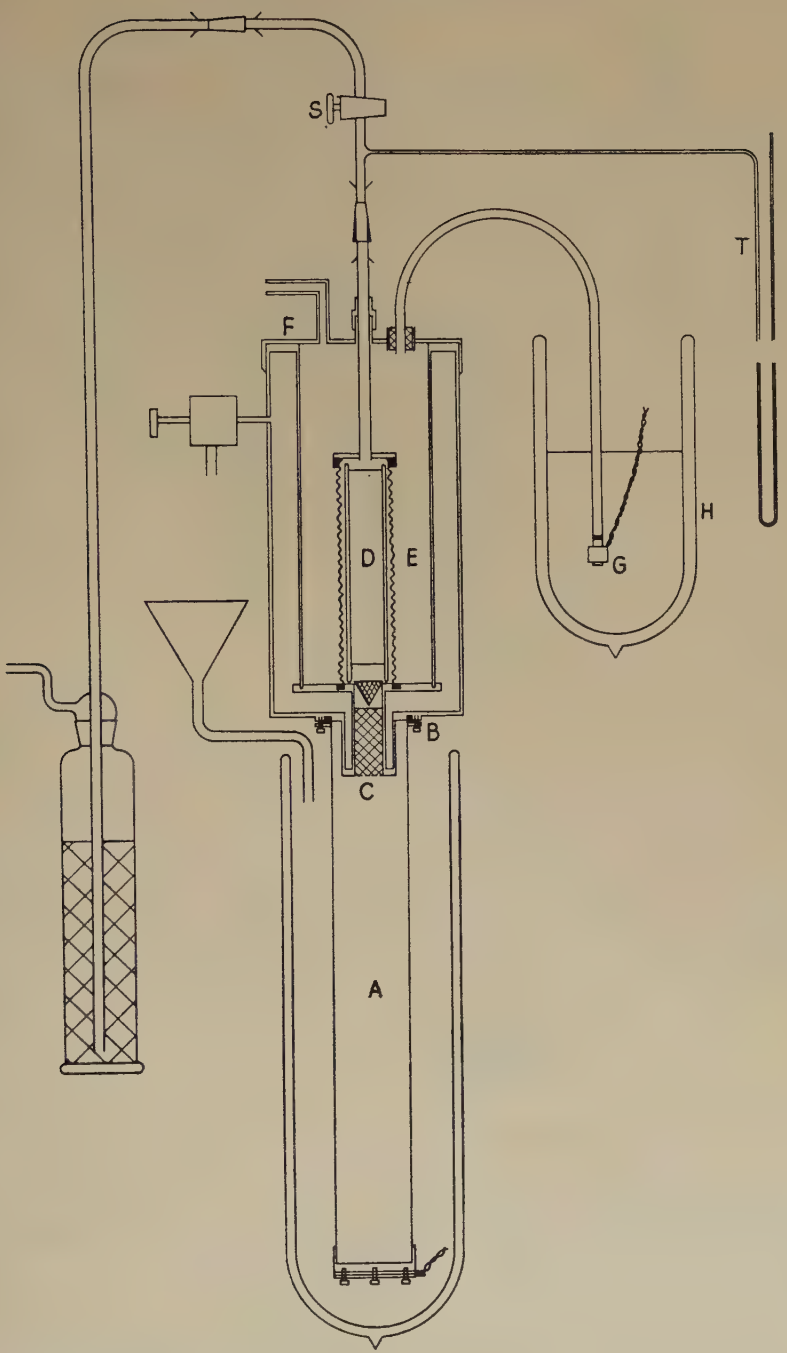

Figure 1. Transportable metal fractionating still for the collection and analysis of atmospheric air samples.

(ii) *Metal apparatus.* In order to facilitate the collection of air samples from atmospheric air which was uncontaminated with industrial waste gases, a metal fractionating still was built in 1954 capable of being transported by van so that samples of air could be collected in a neighbourhood with relatively 'clean' atmospheric air. This still is illustrated diagrammatically in Fig. 1. The boiler A of thin telescopic brass has a capacity of 400 ml and is fitted with a 60 w heater. A is attached to the column and reflux section by a vacuum-tight joint sealed with a lead gasket at B. The short fractionating column at C is filled with stainless steel Dixon Gauze

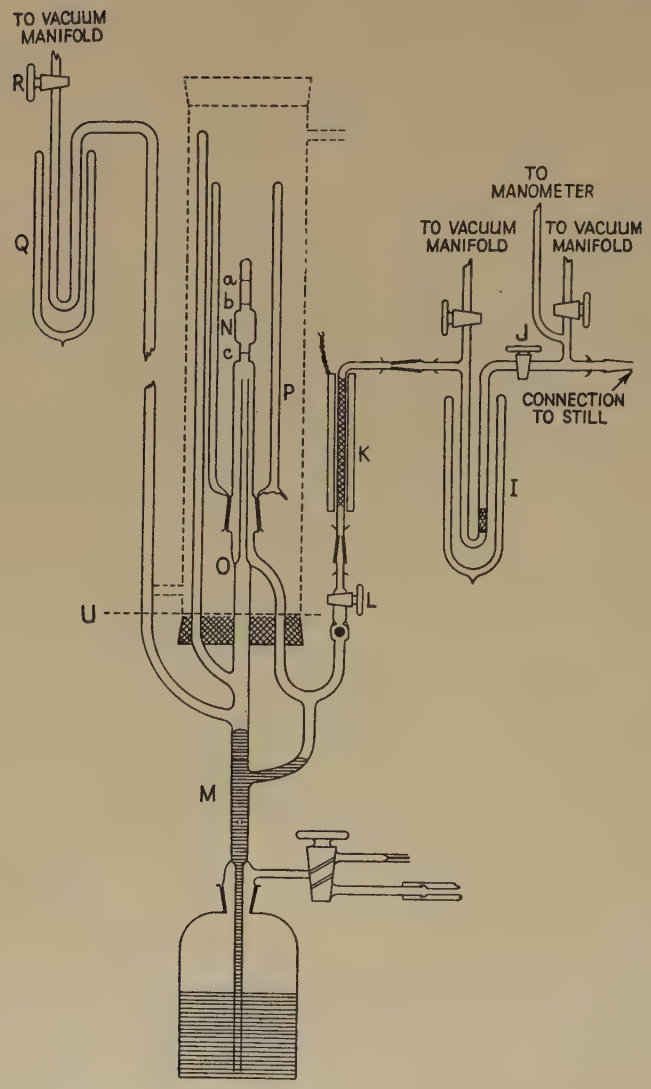

Figure 2. Modified McLeod gauge for measuring small quantities of hydrogen as water vapour.

Ring packing, suitably supported. The reflux section at D is formed from a length of Tombac tubing, the bulk of the dead space being filled with a light brass cylinder supported on legs so as not to impede either liquid or gas flow.

Section D is set in the centre of a vacuum vessel E which is fitted with a lid at F, carrying two entry tubes. Reflux of a liquid-air sample being distilled is accomplished by the pumping of liquid nitrogen held in E, which process results in a drop in the temperature of the nitrogen. As the level of the nitrogen falls, it may be topped up by opening momentarily the electromagnetic valve at G and allowing more nitrogen to be pumped in from the supply held in H.

(*b*) *Copper-oxide tube*

The gases passing from the reflux section of the fractionating column are led by means of hard-glass tubing through the liquid oxygen-cooled U-trap I, which contains a small wad of glass wool, into K, a silica tube containing a quantity of copper oxide in the form of wire (Fig. 2). A small electric heater surrounds K to provide heating to 800-900°C.

(c) *Modified McLeod gauge*

From the heated copper oxide, the water vapour formed from the combustion of hydrogen is passed through stopcock L into the modified McLeod gauge M. The top part of the gauge at N is etched with three lines at a, b and c, the volumes above a mercury-gas interface at these positions being accurately calibrated; the volumes are approximately 0·5, 1·0 and 2·5 ml. At 0 there is an internal seal immediately above which the outside tubing is formed into a cone of standard size to receive the socket of the vacuum vessel P. As is indicated by dotted lines in the diagram the top part of the gauge can be surrounded by a glass jacket. By adjusting the level of mercury in the reservoir, gas passing through stopcock L can be pumped either to vacuum or made to pass through the upper part of the gauge, where, with P in position, filled with liquid oxygen, any water vapour in the gas stream can be condensed out.

From the gauge M, gas can be pumped through the liquid-nitrogen cooled trap Q and stopcock R into the pumping system.

(d) *Gas metering*

After use, the samples of atmospheric air are pumped away through a gas meter attached to the exhaust of the backing pump of the vacuum system.

3. Experimental

(a) *Analysis of air samples*

(i) *Collection of samples – Series A*

In the A series of runs carried out in 1951 using the glass fractionating column, samples were generally taken through a feed line which passed out of the laboratory window, the open end of the feed line being well clear of the ground and away from the side of the building. Usually runs were made when there was a high wind velocity and when the chance of contamination in the air was negligible. Three samples were taken at some distance from the laboratory site in the following way. Four 50-litre carboys, fitted with rubber stoppers and entry tubes with stopcocks, were filled with water and transported from the site to a place with relatively clean atmospheric air. The water was poured away, the stoppers were carefully replaced and the enclosed air samples returned to the laboratory. The condensation of 200 l of air was carried out quite rapidly by surrounding the boiler of the still with liquid nitrogen whilst pumping liquid nitrogen in the reflux section. As the gas was drawn into the still it was passed over a suitable drying agent, either silica gel or anhydrone.

(ii) *Collection of samples – Series B*

The metal still was arranged to be easily detachable from the other part of the apparatus which latter was all of hard glass. To take a sample, the still was evacuated, and the stopcock at S shut; the still was then detached and fitted in a sprung framework for transport. At the sampling location, the vacuum vessel surrounding the boiler and the container around the reflux section were filled with liquid nitrogen. Stopcock S was opened and the air allowed to flow in for approximately 30 min. The pressure on the manometer at T gave an indication of the rate of gas flow and, when necessary, the liquid-nitrogen containers were refilled. As with the glass still, the inflowing air was passed over a suitable dehydrating agent. At the completion of sampling, stopcock S was closed and the still transported to the laboratory. Care was taken to ensure that there was no contamination, e.g., from engine exhaust gases, etc.

(iii) *Processing of samples*

A gas sample in the still was allowed to fractionate at total reflux for a period of 5 min and then a slow flow of gas was drawn off through the trap I, the heated copper oxide in K, the liquid oxygen-cooled trap on the gauge M, the liquid-nitrogen cooled trap Q and the pumping system into the gas meter. The flow of gas was throttled at stopcocks J, L and R which were carefully adjusted so that the pressure in the gauge M was between two and four cm of mercury and the

flow rate was about 12 l/hr. At this flow rate approximately 2 l were withdrawn. Stopcock S was then shut and the gas in the whole of the apparatus beyond this point slowly pumped out. The apparatus was refilled by reopening S which was again closed and the gas slowly pumped away. This procedure was repeated. Eventually, after the bulk of the gas beyond S had been pumped away by careful opening of stopcocks J and L, stopcock L was closed and the gauge M pumped to a hard vacuum.

(iv) *Measurements of water in gauge*

A vacuum of 10^{-4} mm Hg or better, having been obtained in the gauge, the mercury level was raised to the point marked U when the trap P was removed. The mercury was then allowed to fill the gauge completely. The glass jacket was fitted and the gauge head was steamed. After temperature equilibrium had been established, the mercury was slowly lowered to point U, to release any bubbles of water which had been trapped in the lower part of the gauge, and then raised to an appropriate calibration mark. The pressure of water vapour was measured using a cathetometer.

To check that the gas being measured was water vapour, the steam jacket was removed, the gauge allowed to cool down and a small ice cap placed on the top of the calibrated section. Since the pressure of the water vapour above ice is 4·58 mm, any excess above this figure revealed the presence of other gases and where necessary corrections were applied to the results. Alternatively, the water was recondensed, and a second pump-out was given.

(*b*) *Check of method*

(i) *Preparation of standard samples – Series A*

Samples for checking the method were prepared by filling the calibrated bore of a vacuum stopcock with hydrogen at a known pressure and temperature. The glass still was then charged with a quantity of nitrogen by placing a funnel in a vessel containing liquid nitrogen and allowing this gas to be drawn into the still. The sample of hydrogen contained in the stopcock bore was placed in the feed line so that the hydrogen was carried into the still with the nitrogen. Only a few litres of nitrogen were condensed and the run then proceeded as for a normal air sample.

(ii) *Preparation of standard samples – Series B*

A more elaborate method of preparing check samples was developed in which a standard gas sample of nitrogen containing 1 per cent of hydrogen was prepared from which smaller aliquots were taken in an accurated calibrated gas pipette. A small amount of the 1 per cent H_2-N_2 mixture was either fed into the metal still with a larger quantity of pure nitrogen and the normal scheme of analysis followed or the nitrogen stream containing the added hydrogen could be drawn directly through the traps and heated copper oxide into the gauge.

4. Results

(*a*) *A Series: Runs carried out in 1951 with glass still*

The results for this series are tabulated in Table 1. Calibrations carried out between 8 Feb. 1951 and 14 Feb. 1951 gave an almost constant loss of 0·02 ml N.T.P., and the results of runs carried out between 26 Jan. 1951 and 26 Feb. 1951 have been corrected accordingly.

The series of calibrations made between 7 June 1951 and 12 June 1951 showed more scatter but no constant error, so the two runs of 13 June 1951 are uncorrected.

The mean for the A series of runs gives the hydrogen content for atmospheric air at ground level to be 0·74 parts per million.

(*b*) *B Series: Runs carried out in 1954 with transportable still*

Table 2 gives the results of the B series, all samples having been collected on the N. Berkshire Downs at 14.30 hr.

TABLE 1. HYDROGEN CONTENT OF ATMOSPHERIC AIR AT GROUND LEVEL : A SERIES OF EXPERIMENTS

Date and time 1951	Atm. press (cm Hg)	Wind speed and direction (ft/sec)	Result (p.p.m.)	Corrected result (p.p.m.)
–	74·6	–	0·33	0·44
–	74·8	–	0·45	0·56
26 Jan. a.m.	74·4	8 W	1·20	1·29
7 Feb. a.m.	73·95	13 WSW	0·14	0·25
7 Feb. a.m.	73·95	13 WSW	0·54	0·65
8 Feb. a.m.	73·2	23 SSW	0·74	0·85
12 Feb. a.m.	74·6	12 S	1·88	1·99
16 Feb. a.m.	74·1	26 S	1·12	1·23
19 Feb. a.m.	73·6	24 WSW	0·36	0·48
19 Feb. p.m.	73·7	24 WSW	0·02	0·13
19 Feb. p.m.	74·0	20 SW	0·41	0·52
21 Feb. a.m.	73·3	20 SW	0·45	0·56
22 Feb. a.m.	74·1	19 W	0·93	1·05
23 Feb. a.m.	74·7	13 WSW	0·56	0·68
26 Feb. a.m.	75·0	16 SW	0·48	0·59
13 June a.m.	75·5	15 SW	0·76	(0·76)
13 June a.m.	75·5	15 SW	1·06	(1·06)
			mean = 0·74	

* These samples collected on N. Berkshire Downs; all other samples collected at laboratory.

TABLE 2. HYDROGEN CONTENT OF ATMOSPHERIC AIR AT GROUND LEVEL : B SERIES OF EXPERIMENTS

Date 1954	Atm. press (cm Hg)	Wind speed and direction (ft/sec)	Result (p.p.m.)
17 Aug.	75·3	15 S	0·37
18 Aug.	74·8	15 SW	0·43
20 Aug.	75·1	12 NW	0·42
23 Aug.	74·9	14 NW	0·44
24 Aug.	74·5	17 NW	0·62
27 Aug.	75·7	10 SW	0·51
30 Aug.	75·8	17 WSW	0·50
31 Aug.	75·8	9 SW	1·0
1 Sept.	75·3	15 S	0·79
3 Sept.	75·4	20 SW	0·43
16 Sept.	74·3	28 W	0·38
			mean = 0·51

Calibrations, carried out between 16 Aug. 1954 and 15 Sept. 1954, show a wide scatter but the majority of the runs are within ± 10 per cent whilst the others are within ± 20 per cent of the correct figure. The average error over twelve calibration runs is − 0·002 ml N.T.P. which represents for a normal run on atmospheric air a correction of only 0·01 parts per million.

The mean of the analyses on the atmosphere carried out in this series gives a hydrogen content for the atmosphere at ground level of 0·51 parts per million. Combining the results for both series an average hydrogen content of 0·62 parts per million is obtained.

5. DISCUSSION

Although the analytical method employed has not proved to be very accurate and generally results are up to 10 per cent in error, it does seem from the 26 analyses made that the hydrogen content of the atmosphere at ground level is a varying quantity depending on weather conditions prevailing. This should be investigated by taking more than one sample at a time so that any unusual results can be checked.

Acknowledgment

The authors wish to acknowledge the help given by Mr. P. Wharton with some of the experimental work.

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551.574.1 : 551.508.91

Some recent airborne measurements of freezing nuclei over southern England

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(Manuscript received 12 June 1957)

Summary

Daily measurements of freezing-nucleus concentrations at 10,000 ft over southern England during January 1956 revealed no well-marked peaks. The results also indicated, however, that the earth's surface was not the main source of freezing nuclei.

1. Introduction

Bowen (1953) suggested that peaks of rainfall occur about 30 days after meteor showers. A possible connecting mechanism is a temporary increase of freezing nuclei on given dates as the meteoric particles fall into the troposphere. Smith, Kassander and Twomey (1956) have reported measurements in 1955 showing peaks of freezing-nucleus concentration on the expected dates in January, i.e., the 13th, 23rd and 30th, which correspond to the rainfall peaks given by Bowen. Similar measurements were made on 25 days of January 1956 at about 10,000 ft over southern England, using a Hastings aircraft of the Meteorological Research Flight, and these are described below.

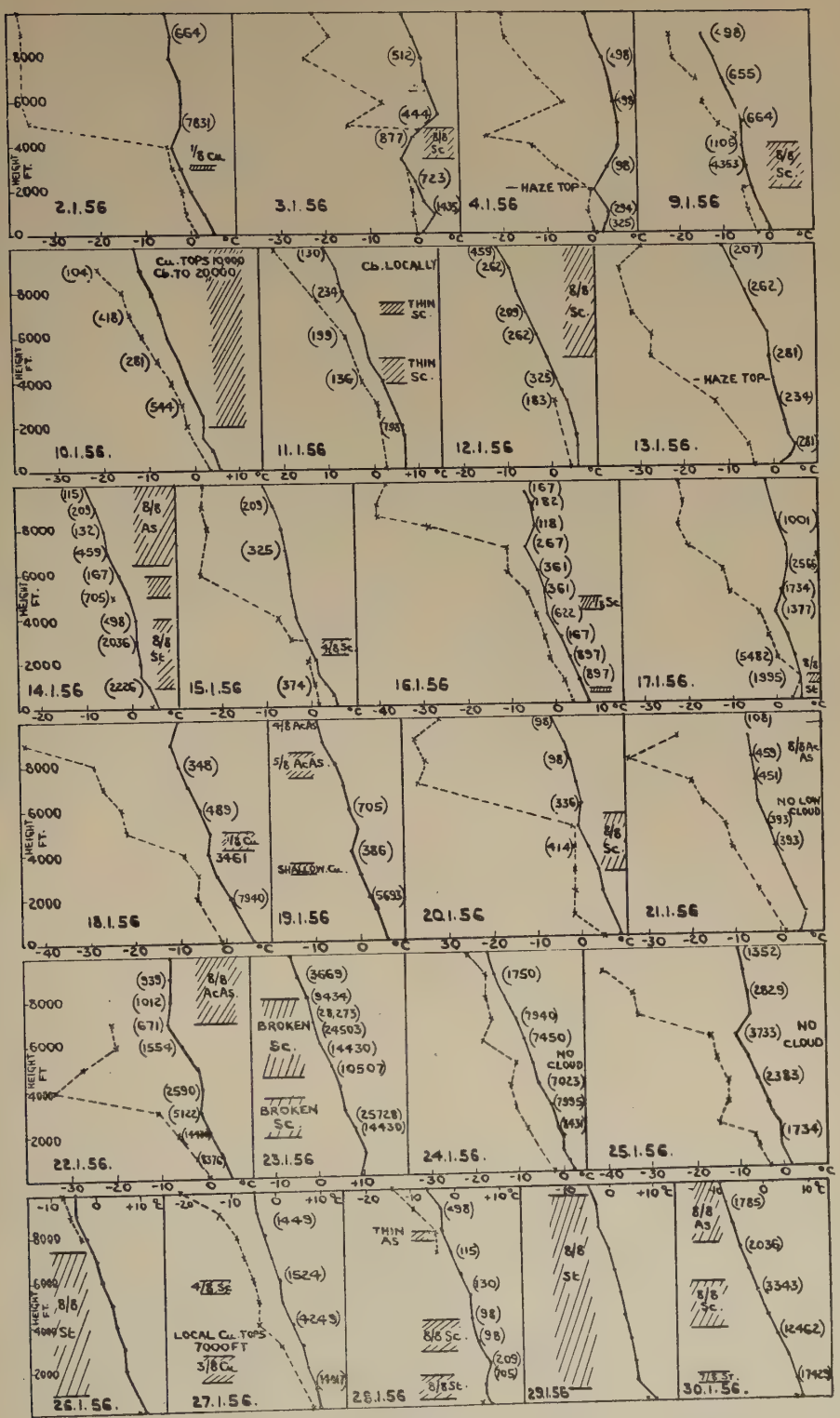

Figure 1. Daily aircraft ascents.

•———• temperature (°C); ×– – –× frost point (°C); ( ) condensation nuclei/cm³; ////// cloud.

2. Instrumentation and flight measurements

A cylindrical cloud chamber, 50 cm deep and 30 cm diameter, rather similar in construction to that described by Smith and Heffernan (1954) was used in these flights. After flushing out the chamber it was sealed and the aircraft remained at the same height for about 30 min while the sample was examined. The temperature at the centre of the chamber was measured as it was cooled continuously to about − 30°C. Steam was injected as required and the interior was illuminated by a parallel beam from a 250 w lamp. The observed illuminated volume was about 0·5 litres. Any ice crystals which formed could easily be distinguished as they sparkled in contrast with the dull opaque background of the droplet cloud. The number of crystals per unit volume at any temperature was calculated from subjective estimates of the apparent spacing between the crystals. In addition to the freezing-nucleus counts the outside air temperatures and humidities were measured at the sampling level and also during the ascent and descent. Condensation-nucleus counts were made as often as possible with a photo-electric counter (Pollak and Murphy 1952) and cloud conditions were noted. The observations below the sampling height are shown in Fig. 1.

3. The daily variation of freezing nuclei over southern England in January 1956

Fig. 2 shows a plot of the temperature at which observed freezing-nucleus concentrations of 1/litre to 10^4/litre occurred on each day of observation. The temperatures at which the first crystal was observed are also shown. Included for comparison is a copy of the results of Smith, Kassander and Twomey (1956) for Sydney, Australia, in January 1955 (Fig. 2). In contrast with the well-marked peaks of their curve the present results do not exhibit any similar variation. The temperature for the 1/litre concentration is always between − 20°C and − 25°C while the day-to-day variation at the 10^2, 10^3 and 10^4/litre level is no more than would be expected from random errors of measurement. No conclusions have been drawn from the temperatures at which the first crystal was seen, these being regarded as very doubtful observations. At the 0·1/litre level it is possible that there was a peak on 28 Jan. The mean temperature for the Meteorological Research Flight's readings is several °C higher than that of the Australian measurements. It is thought that this can be accounted for by differences in chamber design and reading technique, e.g., rates of cooling and position in the chamber at which the temperature measurement

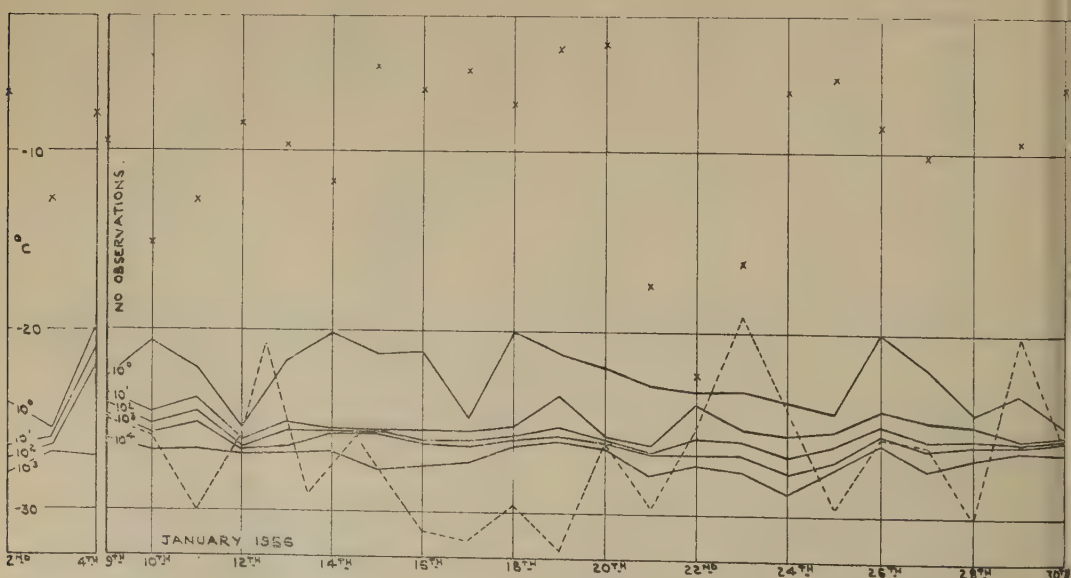

Figure 2. Daily variation of freezing nuclei (10^0, 10^1, 10^2, 10^3 and 10^4/litre).

——— one sample every day at 10,000 ft over southern England; – – – – similar measurements in Australia, Jan. 1955 of 1/litre only; × temperature at which ice crystals were first observed.

was taken. Considering the spectrum as a whole the day of maximum concentrations was 4 Jan., although of course it is possible that there may have been a peak during the period from 5-8 Jan. when no flying was possible.

It appears therefore that this set of measurements gives no support to the hypothesis of Bowen, in that there is no relatively large concentration at the 1/litre level, or above, evident on the expected dates. On the other hand, these measurements are not necessarily evidence against it as they give very little information on concentrations of freezing nuclei between 0°C and — 15°C, the temperature range within which their possible importance in affecting rainfall is the greatest. It appears likely that freezing-nucleus concentrations of 0·1/litre or even smaller may be important in this region. Owing to the limitations of the apparatus, little weight can be attached to the observations at these lower concentrations, but, such as they are, they likewise fail to support the hypothesis.

4. The spectra and source of freezing nuclei

Examples of typical spectra (crystal concentration plotted against temperature) are shown in Fig. 3. In every case the numbers of crystals increase rapidly below — 25°C and it is not obvious whether this is due to an increasing number of nuclei or to spontaneous freezing. On the other hand, although a few crystals were occasionally seen above — 15°C no concentrations of greater than 1/litre were ever obtained above — 18°C. Between — 18°C and — 25°C there is a marked variation in the spectrum shape and nucleus concentration between different occasions and an attempt was made to discover whether this variation could be correlated with other meteorological parameters. On a few occasions when samples were taken at altitude in a beam of silver-iodide particles emitted from the surface, the spectra assumed a form rather similar to that shown for 4 Jan. 1956. It appears that with this apparatus a concentration of about 10/litre at — 10°C to — 15°C is desirable before the silver iodide can be detected in the free atmosphere with confidence.

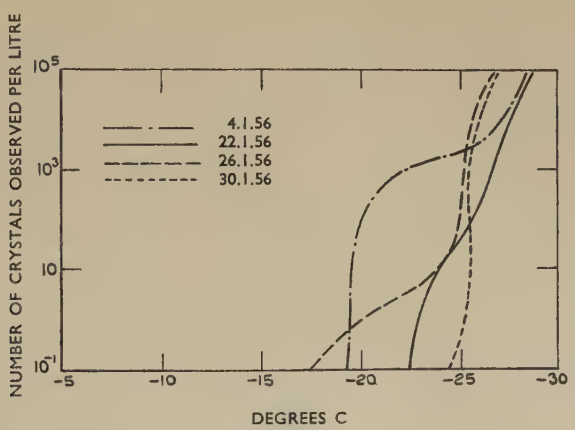

Figure 3. Typical spectra of freezing nuclei at 10,000 ft.

The distribution of condensation nuclei with height shown in the measurements of Fig. 1 suggests that in the main they are produced at the earth's surface and decrease with height. The details of the distribution will probably depend mainly on the location of the different sources and also the vertical stability. If freezing nuclei were also produced at the surface it might be expected that their concentration at 10,000 ft would be high when that of the condensation nuclei is high. This possibility was examined by studying the stability and condensation-nucleus distribution with height on each occasion and the results are shown in Table 1. Freezing-nulceus concentrations at — 20°C and — 25°C are classified as high, moderate or low, and divided into cases of high or low probability that the nuclei could have been readily transported upwards from the earth's surface.

The evidence is that the higher concentrations tended to occur when the vertical transport was unlikely. Furthermore, plots of the daily variation at 10,000 ft of condensation-nucleus concentrations and freezing-nucleus concentrations at — 20°C and — 25°C show little similarity.

TABLE 1. NUMBER OF CASES OF HIGH, MODERATE AND LOW CONCENTRATIONS OF FREEZING NUCLEI AT − 20°C AND − 25°C IN WHICH TRANSPORT UPWARDS FROM THE EARTH'S SURFACE WAS LIKELY

Transport upwards from earth's surface	Freezing-nucleus concentration at − 20°C			at − 25° C		
	High > 1/litre	Moderate 0·1-1·0/litre	Low < 0·1/litre	High > 100/litre	Moderate 10-100/litre	Low < 10/litre
likely	1	4	9	2	3	9
unlikely	2	4	4	1	3	6

In a subsidiary investigation several flights were made in which freezing-nucleus spectra were measured above and below a well-marked temperature inversion. Those taken above the inversion tended always to give higher temperatures of complete glaciation than those below. The mean values which are shown in Fig. 4 indicated a mean temperature difference of about 4°C – a result very similar to that reported by Smith and Heffernan (1954).

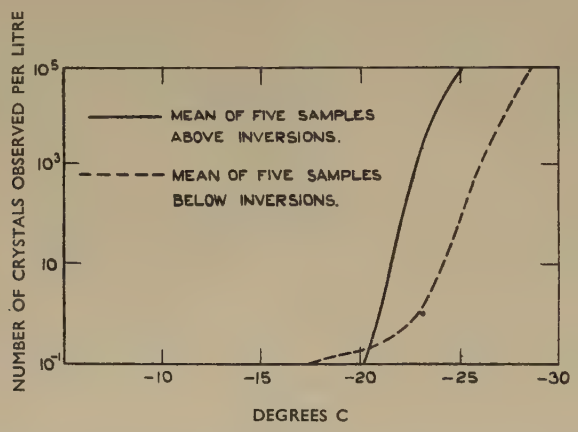

Figure 4. Freezing-nucleus spectra above and below temperature inversions.

Attempts were also made to correlate freezing-nucleus concentration with wind direction and air mass but no dependence on these factors was discovered. Days of rainfall appeared on the whole to have lower concentrations than those found on fine days. Finally a few flights were made in which samples were taken at distances a hundred miles or so apart. These showed changes rather similar in magnitude to those found from day to day.

5. Discussion

When considering these results it should be borne in mind that the cloud-chamber technique both as used on the surface and in the air is crude and measurements based on it are semi-quantitative. This is borne out when one compares previously reported results from different investigators (aufm Kampe and Weickmann 1951). Differences of up to several orders of magnitude have been given for the freezing-nucleus concentrations at a given temperature. Some of the limitations of the airborne equipment have been mentioned by Smith and Heffernan. These include bulk, long operating time, size limitations, subjectivity of measurements, uncontrolled supply of moisture, difficulty of temperature measurement and neglect of possible delay in activation time. There is obviously much scope for improvement in the apparatus both for surface and airborne work.

The results of the present investigation have not supported Bowen's hypothesis in that they have not indicated any preferred dates for the occurrence of large concentrations or higher activation temperatures of freezing nuclei. On the other hand, in common with the results of Smith and Heffernan, they suggest that the earth's surface is not the main source of freezing nuclei, although more measurements are needed before this can be asserted with confidence. The measured temperatures at which considerable freezing took place in the cloud chamber were 9° to 10°C lower than those at which glaciation takes place in clouds. This may be an instrumental effect, but on the other hand it suggests that considerably lower concentrations of freezing nuclei – perhaps down to $1/m^3$ – will have to be measured in order to obtain further evidence of the mechanics of cloud glaciation.

Acknowledgment

Acknowledgment is made to the Director-General of the Meteorological Office for permission to publish this paper.

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551.557.2 (99) : 551.557.36

Upper winds over an Antarctic station

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(Manuscript received 24 May 1957)

1. Introduction

A preliminary note on the upper-air temperatures over Maudheim, the base of the Norwegian-British-Swedish Antarctic Expedition 1949-1952, has already been published by Nils Jorgen Schumacher (1955). This note is intended to throw further light on some points raised in Schumacher's paper and to outline the general pattern of upper winds over Maudheim. Owing to the use of a radio theodolite, the data gave a representative picture of the wind structure over an Antarctic station without the excessive bias of previous observations which were limited to periods of clear weather. The detailed report and discussion of aerological observations made at Maudheim will be published shortly by the Norwegian Polar Institute.

2. Method

Upper winds were determined by means of radio-theodolite (Rawin SCR 658) measurements of the bearing and elevation of a 400 mc transmitter carried on the same balloon as the separate radiosonde transmitter. The height of the balloon from the radiosonde data was used in order to plot the position of the balloon on a Moltschanoff plotting disc. When radiosonde heights were not available, estimated heights based on ascents under similar conditions were used. About half of the data were evaluated at the time of observation, and the remainder were completed after returning to Europe. In addition all corrections have been revised and data replotted as necessary. Wind velocities have normally been averaged over two-minute intervals below about 10 km height and over four-minute intervals above this level. The winds at standard pressure levels have been analysed by means of punched cards.

The values of the standard vector deviation (σ) in Table 1 have been calculated directly without introducing any approximations from

$$\sigma^2 = \frac{V^2}{N} - \mathbf{V}_R^2$$

where V is the wind speed for a given observation and $\mathbf{V}_R$ the resultant mean vector wind speed for the number (N) of observations considered.

The mean shear vectors between certain standard levels have been evaluated by machine for each individual ascent, and these results were then used to obtain mean values shown in Table 2.

3. Data

Upper-wind measurements were made once daily between 1 April 1950 and 10 January 1952 with certain exceptions. On 22 days high winds prevented a successful release of the balloon. Gaps also occurred in May 1950 during reorganization of balloon filling and launching techniques and in February, March 1951 owing to the loss of the Rawin operators then at Maudheim in an accident. Five ascents made during the latter half of March 1950 have also been included.

The mean winds over Maudheim are shown in Table 1, the figures being the mean values of all observations. It is not considered practicable to correct for days when no observations were made, so the figures are slightly biased against the days with high surface winds. The period from 23 February to 25 March 1951 is an exception to this, as for one month after the accident to the Rawin operators, upper winds were measured by visual theodolite only, so the bias for this period is against cloudy weather.

TABLE 1. Direction and speed of upper winds, Maudheim 71°03 S 10°56 W

Level (mb)	Surface	850	700	500	300	200	100	50
				Mean vector wind (tens of degrees – kt)				
Dec.–Feb.	11 – 05	09 – 03	08 – 03	16 – 02	22 – 06	23 – 07	22 – 05	14 – 03
Mar.–May	12 – 07	12 – 02	22 – 07	23 – 18	23 – 31	24 – 28	25 – 32	24 – 33
June-Aug.	09 – 10	07 – 07	22 – 04	23 – 12	24 – 22	25 – 25	25 – 36	25 – 47
Sept.-Nov.	11 – 08	10 – 05	19 – 04	22 – 13	23 – 21	23 – 23	25 – 28	26 – 32
				Standard vector deviation (kt)				
Dec.-Feb.	12	16	14	19	29	16	10	7
Mar.-May	13	19	24	36	49	29	31	26
June-Aug.	15	25	25	33	42	37	28	24
Sept.-Nov.	14	21	23	26	39	34	28	32
				Mean scalar speed (kt)				
Dec.-Feb.	11	13	13	17	24	14	9	6
Mar.-May	12	16	21	31	44	34	37	36
June-Aug.	15	22	21	30	40	36	41	47
Sept.-Nov.	14	18	18	25	37	34	35	38
				Number of observations				
Dec.-Feb.	121	118	118	117	115	115	104	54
Mar.-May	130	128	127	126	123	117	96	33
June-Aug.	182	179	178	180	179	176	92	2
Sept.-Nov.	175	172	173	172	170	169	151	73

The general pattern of the results shows a predominantly easterly wind at low levels. At the surface a frequency analysis shows that winds tend to blow from E to NE and from S to SW most frequently, with all gales coming from the former direction. Tables 1 and 2 show the tendency of winds to swing to SW and W with increasing height, while a frequency analysis shows that the spread of wind velocities about the mean value approaches a normal distribution at high levels. Wind speeds in the troposphere tend to be highest a little below the tropopause, then, after some decrease around the tropopause, the westerly winds tend to increase with height in the stratosphere, except for summer months. The light stratospheric winds during summer months show an easterly component at the 50 mb level.

TABLE 2. MEAN WIND SHEAR VECTORS OVER MAUDHEIM (TENS OF DEGREES – kt)

Levels (mb)	700-850	500-700	300-500	100-300	50-100
Dec.-Feb.	0	23 – 04	24 – 05	11 – 01	08 – 06
Mar.-May	24 – 08	23 – 11	23 – 13	31 – 09	28 – 07
June-Aug.	24 – 10	24 – 08	24 – 10	29 – 20	26 – 17
Sept.-Nov.	23 – 07	23 – 09	24 – 09	29 – 12	30 – 08
			Number of observations		
Dec.-Feb.	117	115	114	103	54
Mar.-May	126	124	123	95	32
June-Aug.	178	178	179	92	2
Sept.-Nov.	172	172	169	155	73

The variation of the standard vector deviation with height is similar to that found in other latitudes by Brooks, Durst, Carruthers, Dewar and Sawyer (1950), as it increases to a maximum value a little below the tropopause, then after decreasing at first it remains roughly constant above 140 mb.

The mean wind shear vectors in the troposphere are generally from 230° to 240° but swing significantly to a direction of about 290° in the stratosphere, except in the summer months. In this case the easterly mean wind shear vector in the stratosphere is quite definite, indicating that the horizontal temperature gradient is the reverse of normal, with warmer air lying to the south.

4. ACCURACY

A discussion of the accuracy of Rawin measurements under good conditions has been given by Kirkman and le Bedda (1948). Owing to constantly changing snow-drifts around the Rawin hut, additional errors were frequently introduced, while at other times the snow-drifts improved the accuracy of measurements by cutting off surface reflections of the 400 mc waves. This effect has since been discussed in detail by Hey and Parsons (1956) who showed how wire screens, similar in action to our fortuitous snow-drifts, can be used to improve the accuracy of radar elevation measurements at low angles. The system of allowing for the effect of snow-drifts and the resultant accuracy of elevation and bearing measurements is being discussed in the detailed report on upper winds. The errors of observation are not sufficient to obscure the main features of wind structure presented in this note.

Another difficulty in analysing upper-wind observations is to avoid bias of the results for upper levels against conditions of strong upper winds due to loss of contact with the balloon at long range. In this respect the Rawin method is superior to the direct radar technique, as it is possible to follow the balloon to greater distances than 100 km. A check on this type of error can be obtained by comparing the mean shear vector in Table 2, based on individual ascents, with the shear vector obtained by taking the difference between the mean vector winds at standard levels shown in Table 1. A bias of the mean wind data in Table 1 against strong upper winds will then be shown by a difference in the shear vector obtained by the two methods. However, the resultant shear vectors agree satisfactorily to within 1·5 kt up to the 100 mb level, except for the 100–300 mb shear vector during the winter months (June-August) when the difference is 4 kt, and for the 50–100 mb range in autumn and winter when the differences are about 8 kt. Schumacher (private communication) has independently concluded that the mean values of winds above the 300 mb level during winter are the least reliable portion of Table 1, probably owing to bias against strong southerly winds when stratospheric temperatures were low. The cause of the bias may be due to the bursting of the balloon at lower levels under these conditions, rather than to actual failure of the Rawin instrument itself.

5. STRATOSPHERIC TEMPERATURES AND MERIDIONAL EXCHANGE

In the preliminary note on temperature analysis, Schumacher has suggested that a radiative effect exists which could explain the much lower stratospheric temperatures found over Antarctica than over the Arctic during winter months. On the other hand Court (1942) and Ruben (1953) conclude that a stronger meridional exchange of air in the northern hemisphere is a more probable

reason for the warmer stratosphere in the Arctic during winter months. A comparison of upper winds over Maudheim with corresponding winds at 70°N has therefore been made in Table 3 to see if any factors of general significance emerge.

As upper-wind summaries for the Arctic are nearly as scarce as for the Antarctic, the charts published by Brooks *et al.* (1950) in ' Upper winds over the world ' have been used to provide Arctic data. Figures for Tromsö, 70°N, 20°E, have been taken from the charts as this is a somewhat similar location to Maudheim, to the east of a ridge of high pressure at upper levels. In addition maximum and minimum values of standard vector deviation along the latitude 70°N and the mean vector wind with the maximum North-South component across this latitude are tabulated on a seasonal basis in Table 3.

TABLE 3. COMPARISON OF UPPER WINDS, MAUDHEIM AND 70°N

Level mb	Mean vector winds (tens of degrees – kt) Maudheim DD – ff	Tromsö DD – ff	Max N-S component on 70°N DD – ff	Standard vector deviation (kt) Maudheim	Tromsö	70°N Max	70°N Min
			Summer				
700	08 – 03	23 – 08	32 – 30	14	20	23	12
500	16 – 02	22 – 14	21 – 17	19	27	27	18
300	22 – 06	25 – 22	25 – 23	29	36	36	25
200	23 – 07	24 – 25	31 – 34	16	28	28	24
130	23 – 06	26 – 20	30 – 26	11	20	21	17
			Autumn				
700	22 – 07	24 – 13	31 – 19	24	24	26	14
500	23 – 18	25 – 16	33 – 24	36	30	32	20
300	23 – 31	24 – 27	32 – 32	49	40	41	25
200	24 – 28	26 – 30	25 – 31	29	37	37	22
130	24 – 30	26 – 19	25 – 33	32	24	25	18
			Winter				
700	22 – 04	25 – 10	02 – 15	25	24	27	13
500	23 – 12	27 – 15	18 – 20	33	30	35	21
300	24 – 22	27 – 27	24 – 47	42	41	43	30
200	25 – 25	28 – 33	31 – 47	37	34	35	26
130	25 – 31	30 – 36	23 – 48	29	25	28	20
			Spring				
700	19 – 04	29 – 11	30 – 17	23	26	27	16
500	22 – 13	31 – 13	29 – 26	26	31	36	18
300	23 – 21	30 – 21	23 – 32	39	40	40	28
200	23 – 23	30 – 28	31 – 26	34	33	35	25
130	24 – 25	31 – 25	30 – 41	28	25	27	19

During summer months, the movement of air in the Antarctic stratosphere appears to be less than over the Arctic, since the standard vector deviation over Maudheim is lower than the minimum value at 70°N, and the N-S component at Maudheim is small. During the autumn, however, the figures indicate that meridional movement over Maudheim is of the same order as, or greater than, the maximum meridional movement of the stratosphere over 70°N. During spring the stratospheric movement over Maudheim and 70°N appear to be of the same order of magnitude. On the other hand the N-S component of movement in the stratosphere during winter is much lower over Maudheim than the maximum across 70°N, but not very different from values over Tromsö.

Although the accuracy of figures from both 70°N and Maudheim is not high, the comparison does indicate some significant points. The small meridional exchange of air over the Antarctic during summer months might provide the explanation of slightly higher temperatures in the stratosphere over the Antarctic than over the Arctic at this season. Unfortunately the corresponding data for winter months are inconclusive, partly owing to the probable bias of the Maudheim results at this season. Table 3 indicates that meridional exchange in the stratosphere

over Maudheim may be a little lower than the mean at 70°N, but it is well within the extreme limits over the Arctic. The autumn figures are, however, of much more significance in assessing the influence of meridional exchange on the low stratospheric temperatures over the Antarctic during winter months. In spite of inaccuracies and having figures from only one Antarctic locality for comparison with 70°N, it appears unlikely that meridional exchange in the Antarctic stratosphere is much lower than in the Arctic at this season. Nevertheless Schumacher's paper shows that the temperature at the 100 mb level falls three times as rapidly over Maudheim as over Arctic Bay during autumn months. The major cooling effect over the Antarctic during autumn is not therefore due to small meridional exchange of air. Consequently it is most unlikely that differences in meridional exchange explain the lower temperatures of the Antarctic stratosphere during winter, and radiative effects may well provide the explanation.

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551.511.6 : 551.551.21

Wind profiles and the shearing stress – an anomaly resolved

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Summary

Reported failures of a generalized wind-profile equation to predict correctly the shearing stress under certain conditions are found to be associated with low wind speeds close to the surface. The increase of drag coefficient of the surface-roughness elements (e.g., grass blades) at low Reynolds numbers is the source of these discrepancies.

1. Introduction

In a study of the vertical profiles of wind over various level surfaces (Deacon 1949, 1953) the well-known logarithmic relationship for aerodynamically rough flow under neutral conditions was generalized to give the equation

$$u = \frac{(\tau/\rho)^{\frac{1}{2}}}{k\,(1-\beta)} \left\{ \left(\frac{z}{z_0} \right)^{1-\beta} - 1 \right\} \quad . \quad . \quad . \quad . \quad (1)$$

where u is the wind speed at height z, τ is the shearing stress, ρ the air density, k is the Kármán constant ($\simeq 0{\cdot}40$), z_0 the roughness parameter and β depends on the stability, being > 1 for unstable conditions and < 1 for stable.

It has generally been found that Eq. (1) gives a reasonably satisfactory representation of the form of wind profiles in the lowest few metres but the more stringent test is whether it allows the shearing stress to be correctly predicted from wind observations. Rider (1954) found as a result of direct measurement of τ that the method was satisfactory for unstable conditions but considerably underestimated the shearing stress under stable conditions. It is the purpose of this note to suggest a reason for this failure.

2. Evidence and interpretation

If Eq. (1) were exact then it would be possible, in principle, to evaluate τ simply from the observations of wind speed at several heights. In practice, because of observational error, departures from horizontal uniformity and the approximate nature of the formulation, this procedure is unworkable and it is necessary to adopt another course. This is to obtain a value of z_0 from wind profile observations under neutral conditions and to assume this holds also for the non-neutral conditions. Wind-speed observations at two heights then give the necessary data for the evaluation of β and τ.

The assumption of a z_0 independent of stability appeared reasonable in view of the evidence that quite close to the surface the effects of buoyancy on the flow conditions appear to be small but the matter needs closer examination. A test of this assumption would be provided by measurements of the surface drag together with measurements of the wind speed (u_s) close above the surface-roughness elements. A z_0 independent of stability would correspond to a constant value of $\tau/(\rho u_s^2)$. The lowest level for which Rider (1954) gives wind speeds is 15 cm but this should be close enough to the surface for the present purpose and Fig. 1 has accordingly been prepared from his data. It is apparent that for $u_{15} > 1{\cdot}5$ m/sec there is little evidence of any systematic variation of $\tau/(\rho u_{15}^2)$ with either wind speed or stability. Below 1·5 m/sec, however, there is evidence of a marked rise of $\tau/(\rho u_{15}^2)$ as the wind speed decreases and this applies both to stable and unstable conditions.

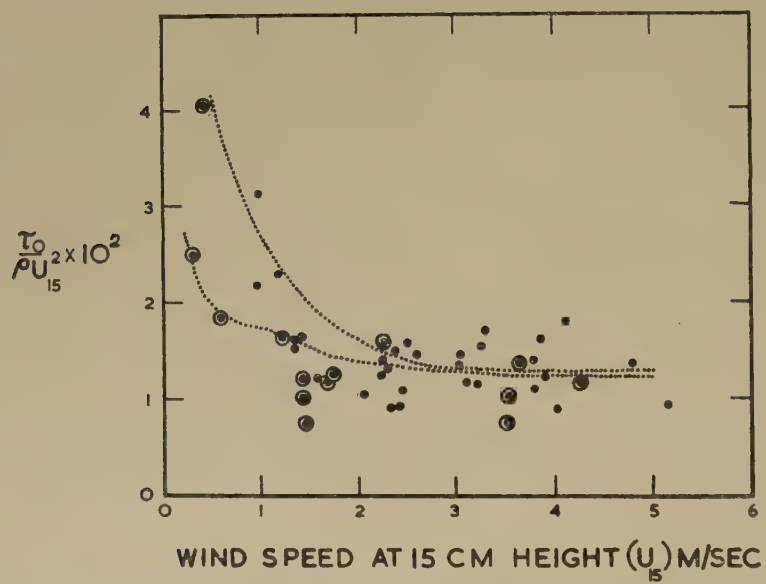

Figure 1. The variation of the drag coefficient of a grass surface with wind speed (Rider's 1954 data). ⊙ Stable stratification. • Unstable stratification. For the significance of the dotted lines see text.

Ellison (1956) has remarked on this feature of Rider's results but he doubts the reality of the effect – he attributes it to experimental error in the measurement of the small forces under these conditions. But, if we accept equality of the transfer coefficients for momentum and vapour, for which there is fairly strong evidence (e.g. Rider 1954, Deacon and Swinbank 1956), then Pasquill's (1949) work on evaporation supports an increase of drag coefficient at low wind speeds. As in Pasquill's experiments the low wind speeds occurred under stable conditions it seemed that the failure of estimates based on Eq. (1) and a constant z_0 was probably an effect of stability. The following considerations suggest it to be rather a result of the low wind speed.

At low Reynolds numbers the turbulent wakes of the surface-roughness elements (grass blades in the studies referred to above) are suppressed in some degree and this results in a rise in drag coefficient at low speeds as is illustrated by the data shown in Fig. 2 for circular cylinders

Goldstein 1938) and an aerofoil (Thom and Swart 1940). In the latter case the curve for a 10° ngle of incidence, for which the data were most complete, is given, but other angles give similar ariations.

To relate these results to Rider's data in Fig. 1 it is assumed that the blades of the short rass behave somewhat similarly to cylinders of about 1 mm diameter or to aerofoils of chord mm. On this basis and also taking the wind speed close above the grass to be $\frac{2}{3}$ of that at 15 cm, he curves of Fig. 2 have been transferred to Fig. 1 as the dotted lines, the upper one being for he aerofoil. The ordinates have been adjusted in scale to coincide with Rider's results at 3 m/sec. The wind speed at 15 cm of 1·5 cm/sec, below which $\tau/(\rho u_{15}^2)$ rises markedly, corresponds on his basis to a Reynolds number of 70 for 1 mm cylinders and of 210 for aerofoils of 3 mm chord. These are, of course, only approximate values in view of the estimates involved. The value of u_s/u_{15} might, for example, be nearer $\frac{1}{2}$ than $\frac{2}{3}$. It is therefore not intended by the comparison of Fig. 1 to convey any very exact correspondence. This would in any case hardly be expected, as with assemblages of roughness elements mutual interference is likely to be appreciable.

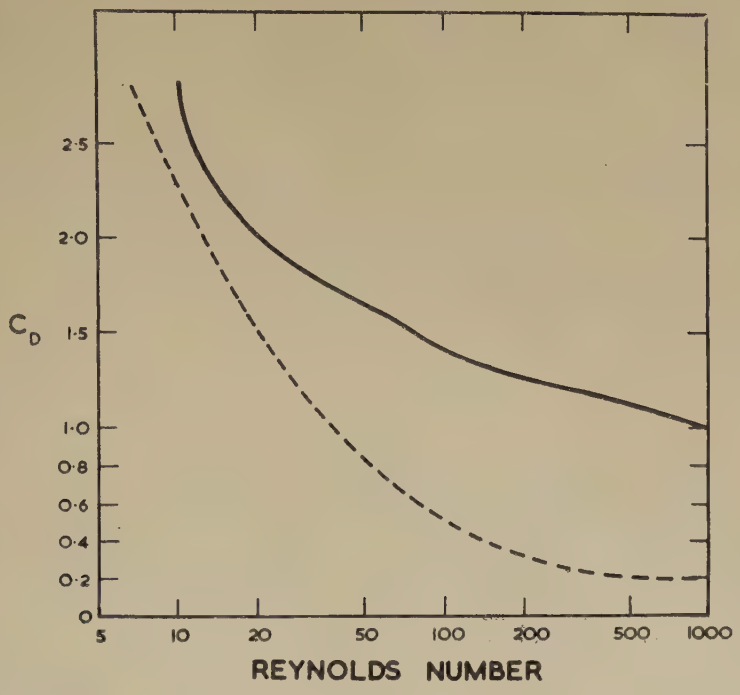

Figure 2. The variation of drag coefficient (C_D) with Reynolds number for a circular cylinder (full line) and for an aerofoil at 10° angle of incidence (broken line).

It is concluded from this comparison that the rise in $\tau/(\rho u_{15}^2)$ at low speeds is a result of the Reynolds number of the flow about the grass blades being too low under these conditions for fully rough flow to exist.

For surfaces roughened with sand grains, as used by Nikuradse in his well-known experiments, fully rough flow occurs at all speeds above those equivalent to

$$\frac{u_* z_0}{\nu} > 2{\cdot}5 \quad . \quad . \quad . \quad . \quad . \quad (2)$$

where $u_* = (\tau/\rho)^{\frac{1}{2}}$, and according to Schlichting (1936) ' It is probable that approximately the same limit also applies to other surfaces.' The above analysis shows, however, that for the grass Rider worked with ($z_0 = 0{\cdot}3$ cm), the corresponding value of $(u_* z_0)/\nu$ is about 40. This supports Schlichting's (1955) later statement that the criterion (2) is only valid for surfaces geometrically similar to Nikuradse's sanded surfaces.

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Exact reflectance under a cardioidal luminance distribution

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Summary

The *cardioidal luminance distribution* $B(\theta, k) = B(\pi/2, k)(1 + k\cos\theta)$, $-1 \leqq k < \infty$, $0 \leqq \theta \leqq \pi/2$ is the general empirical form of the luminance of overcast skies. It has been found empirically (Moon and Spencer 1942) and theoretically (Chandrasekhar 1946) that the value $k = 2$ represents most overcast skies. The purpose of this note is to obtain the exact algebraic expression for the reflectance $r(m, k)$ of a surface (between media of relative indices of refraction 1 and $m \geqq 1$) irradiated by an unpolarized cardioidal radiance distribution. In this way the classical reflectance formula of Walsh (1926) for the case $k = 0$ is generalized, and the approximately-obtained numerical results of Middleton (1952) for the case $k = 2$, $m = 4/3$ are extended. In addition it is found that the function $r(\cdot, k)$ is essentially linear over the range $1{\cdot}2 \leqq m \leqq 1{\cdot}9$ for each fixed value of k in the interval $0{\cdot}5 \leqq k < \infty$, and thus may be represented by the form : $r(m, k) = a(k)\, m + b(k)$. In particular $r(m, 2) = 0{\cdot}141965\, m - 0{\cdot}137709$ which has a maximum deviation of 3·2 per cent from the exact values of $r(m, 2)$ over the range $1{\cdot}2 \leqq m \leqq 1{\cdot}9$.

1. Introduction

The purpose of this note is to give an exact expression for the reflectance of a surface illuminated by a *cardioidal luminance distribution*, i.e., a luminance distribution of the form :

$$B(\theta, k) = B(\pi/2, k)(1 + k\cos\theta), \quad -1 \leqq k < \infty, \quad 0 \leqq \theta \leqq \pi/2 \quad . \qquad (1)$$

The exact expression is used to obtain numerical values of the reflectance for a given set of values of k and the relative index of refraction m. In this way the classical exact expression for the case $k = 0$ given by Walsh (1926) is generalized (Eq. (14)), and certain results obtained by Middleton (1952) are extended (Table 1).

Before proceeding with the details, we observe that Eq. (1) with $k = 2$ is the empirical form for the luminance of an overcast sky, as reported by Moon and Spencer (1942) and recently confirmed by Hopkinson (1954). It is of interest to note that this empirical expression is in good agreement with the corresponding theoretical expression deduced by Chandrasekhar (1946).

2. The reflectance $r(m, k)$

Let S be a surface between two optical media M and M' of relative indices of refraction (for a fixed wavelength λ) 1 and $m \geqq 1$ respectively. Let S be irradiated by an unpolarized radiance distribution coming from M of the form $N(\theta, k) = N(\pi/2, k)(1 + k \cos\theta)$ where θ is measured in radians from the normal to S in M and k is any fixed real number such that $-1 \leqq k < \infty$. Required to find the exact form of the spectral reflectance $r(m, k)$ of S with respect to $N(\cdot, k)$.

At a given point p of the surface S, the spectral reflectance r with respect to a radiance distribution N is defined as

$$r = W/H, \qquad (2)$$

where W is the radiant emittance of, and H is the irradiance on, S at p as induced by N. In the case of the distribution $N(\cdot, k)$, let $W(m, k)$ and $H(k)$ be the associated emittance and irradiance. Clearly

$$W(m, k) = 2\pi \int_0^{\pi/2} N(\theta, k)\, r'(\theta, m) \cos\theta \sin\theta \, d\theta, \qquad (3)$$

and

$$H(k) = 2\pi \int_0^{\pi/2} N(\theta, k) \cos\theta \sin\theta \, d\theta, \qquad (4)$$

where

$$r'(\theta, m) = \tfrac{1}{2}[\sin^2(\theta - \phi)\operatorname{cosec}^2(\theta + \phi) + \tan^2(\theta - \phi)\cot^2(\theta + \phi)] \qquad (5)$$

is the Fresnel reflectance formula in which θ and ϕ are related by Snell's Law : $\sin\theta = m \sin\phi$. Hence

$$r(m, k) = W(m, k)/H(k) = W'(m, k)/H'(k) \qquad (6)$$

where

$$W'(m, k) = W(m, k)/2\pi N(\pi/2, k), \quad H'(k) = H(k)/2\pi N(\pi/2, k) \qquad (7)$$

It follows that

$$H'(k) = \int_0^{\pi/2} (1 + k \cos\theta) \cos\theta \sin\theta \, d\theta = \tfrac{1}{2} + k/3 \qquad (8)$$

Further, let $\theta' = \theta - \phi$, then r' may be written

$$r'(\theta', m) = [2m^2/(m^2 - 1)^2][(\cos\theta' - a)^2 + (\sec\theta' - a)^2] \qquad (9)$$

in which $a = (m^2 + 1)/2m$, and $0 \leqq \theta' \leqq \cos^{-1}(1/m)$. Letting $x = \cos\theta'$, $s = 1/m$, it follows that

$$W'(m, k) = \tfrac{1}{2}\int_1^s [1 + k(x - s)/(1 + s^2 - 2sx)^{\frac{1}{2}}]\{[(x - a)^2 + (1/x - a)^2]/(1 - s^2)^2\} \times \times [(sx - 1)(x - s)/(x - a)^2]\, dx \qquad (10)$$

Clearly $W'(m, k)$ can be written in the form

$$W'(k, m) = A(m) + kB(m) \qquad (11)$$

where

$$A(m) = \tfrac{1}{2}\int_1^s \{[(x - a)^2 + (1/x - a)^2][(sx - 1)(x - s)]/(x - a)^2(1 - s^2)^2\}\, dx \qquad (12)$$

and

$$B(m) = \tfrac{1}{2}\int_1^s \{[(x - a)^2 + (1/x - a)^2][(sx - 1)(x - s)^2]/(x - a)^2(1 - s^2)^2(1 + s^2 - 2sx)^{\frac{1}{2}}\}\, dx \qquad (13)$$

So that

$$r(m, k) = 6[A(m) + kB(m)]/(3 + 2k) \qquad (14)$$

The classical expression considered by Walsh is

$$r(m, 0) = 2A(m) \qquad (15)$$

The evaluation of $A(m)$ is straightforward and yields :

$$2A(m) = \tfrac{1}{2} + (m - 1)(3m + 1)/6(m + 1)^2 - 2m^3(m^2 + 2m - 1)/(m^2 + 1)(m^4 - 1) + [8m^4(m^4 + 1)/(m^2 + 1)(m^4 - 1)^2] \ln m + [m^2(m^2 - 1)^2/(m^2 + 1)^3] \ln[(m - 1)/(m + 1)] \qquad (16)$$

For $k \neq 0$, the new contribution is from the term $B(m)$. The most natural representation of $B(m)$ is :

$$B(m) = [m/2(m^2-1)^2]\{\sum_{i=1}^{4} B_{1i}(m) + m[\sum_{i=1}^{4} B_{2i}(m) - (a^2-1)\sum_{i=1}^{4} B_{3i}(m)]\} \quad (17)$$

where, with the abbreviations

$$p = m-1, \quad q = m^2-1, \quad r = m^2+1, \quad . \quad . \quad . \quad (18)$$

$$B_{11} = (1/8m)[r^3(p - q^{\frac{1}{2}}) - r^2(p^3 - q^{\frac{3}{2}}) + (3r/5)(p^5 - q^{\frac{5}{2}}) - (1/7)(p^7 - q^{\frac{7}{2}})] \quad . \quad (19)$$

$$B_{12} = [(m^2+2)/2m][(-r^2/2)(p - q^{\frac{1}{2}}) + (r/3)(p^3 - q^{\frac{3}{2}}) - (1/10)(p^5 - q^{\frac{5}{2}})] \quad . \quad (20)$$

$$B_{13} = [(2m^2+1)/2m][r(p - q^{\frac{1}{2}}) - (1/3)(p^3 - q^{\frac{3}{2}})] \quad . \quad . \quad . \quad . \quad (21)$$

$$B_{14} = -m(p - q^{\frac{1}{2}}) \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (22)$$

$$B_{21} = (r^2/8m^2)[r(p - q^{\frac{1}{2}}) - (1/3)(p^3 - q^{\frac{3}{2}})] \quad . \quad . \quad . \quad . \quad (23)$$

$$B_{22} = [-r(5m^2+1)/4m^2](p - q^{\frac{1}{2}}) \quad . \quad . \quad . \quad . \quad . \quad (24)$$

$$B_{23} = [(2m^2+1)/r^{\frac{1}{2}}] \ln A, \quad A = (q^{\frac{1}{2}} - r^{\frac{1}{2}})(p + r^{\frac{1}{2}})/(q^{\frac{1}{2}} + r^{\frac{1}{2}})(p - r^{\frac{1}{2}}) \quad . \quad . \quad (25)$$

$$B_{24} = (-m/r)(p - mq^{\frac{1}{2}}) - (m^2/r^{\frac{3}{2}}) \ln A \quad . \quad . \quad . \quad . \quad . \quad (26)$$

$$B_{31} = (r^2/2)[(p^{-1} - q^{-\frac{1}{2}}) - (r/3)(p^{-3} - q^{-\frac{3}{2}})] \quad . \quad . \quad . \quad . \quad (27)$$

$$B_{32} = [r(5m^2+1)/3](p^{-3} - q^{-\frac{3}{2}}) \quad . \quad . \quad . \quad . \quad . \quad . \quad (28)$$

$$B_{33} = [4m^2(2m^2+1)/r][(-2/r)(p^{-1} - q^{-\frac{1}{2}}) - (2/3)(p^{-3} - q^{-\frac{3}{2}}) + (1/r^{\frac{3}{2}}) \ln A] \quad (29)$$

$$B_{34} = (40m^3/r^3)(p^{-1} - q^{-\frac{1}{2}}) + (40m^4/3r^2)(p^{-3} - q^{-\frac{3}{2}}) - (4m^3/r)(p^{-3} - mq^{-\frac{3}{2}}) - (20m^4/r^{\frac{3}{2}}) \ln A \quad . \quad . \quad . \quad . \quad . \quad (30)$$

3. Numerical results

Computations based on the exact expression for $r(m, k)$ over the ranges indicated below were carried out to nine places and rounded to six.

TABLE 1. The reflectance $r(m, k)$

				k			
m	$-1{\cdot}0$	$-0{\cdot}5$	0	$0{\cdot}5$	$1{\cdot}0$	$1{\cdot}5$	$2{\cdot}0$
$1{\cdot}1$	0·054315	0·032448	0·025159	0·021515	0·019328	0·017870	0·016829
$1{\cdot}2$	0·088540	0·055346	0·044281	0·038749	0·035429	0·033216	0·031636
$1{\cdot}3$	0·112699	0·074024	0·061132	0·054686	0·050819	0·048240	0·046399
4/3	0·119641	0·079754	0·066458	0·059811	0·055822	0·053163	0·051264
$1{\cdot}4$	0·132357	0·090698	0·076811	0·069858	0·065702	0·062925	0·060941
$1{\cdot}5$	0·149237	0·106143	0·091778	0·084595	0·080286	0·077413	0·075361
$1{\cdot}6$	0·163601	0·120584	0·106245	0·099076	0·094774	0·091907	0·089858
$1{\cdot}7$	0·177942	0·134726	0·120320	0·113117	0·108796	0·105915	0·103857
$1{\cdot}8$	0·190639	0·148201	0·134055	0·126982	0·122738	0·119909	0·117888
$1{\cdot}9$	0·202559	0·161246	0·147476	0·140590	0·136459	0·133705	0·131737
m	$2{\cdot}5$	$3{\cdot}0$	$3{\cdot}5$	$4{\cdot}0$	$4{\cdot}5$	$5{\cdot}0$	∞
$1{\cdot}1$	0·016048	0·015440	0·014955	0·014557	0·014226	0·013945	0·010581
$1{\cdot}2$	0·030450	0·029528	0·028790	0·028187	0·027684	0·027258	0·022152
$1{\cdot}3$	0·045017	0·043943	0·043084	0·042380	0·041794	0·041299	0·035349
4/3	0·049839	0·048731	0·047845	0·047119	0·046515	0·046004	0·039867
$1{\cdot}4$	0·059453	0·058296	0·057370	0·056613	0·055982	0·055448	0·049039
$1{\cdot}5$	0·073822	0·072625	0·071667	0·070883	0·070230	0·069678	0·063048
$1{\cdot}6$	0·088322	0·087127	0·086171	0·085389	0·084737	0·084186	0·077568
$1{\cdot}7$	0·102313	0·101113	0·100152	0·099367	0·098712	0·098158	0·091509
$1{\cdot}8$	0·116372	0·115193	0·114250	0·113479	0·112836	0·112292	0·105763
$1{\cdot}9$	0·130262	0·129114	0·128196	0·127445	0·126819	0·126290	0·119934

4. Approximate forms for $r(m, k)$

A plot of the numerical results would show that $r(\cdot, k)$ is essentially a linear function for the ranges $1{\cdot}2 \leqq m \leqq 1{\cdot}9$, $\frac{1}{2} \leqq k \leqq \infty$. Hence over these ranges $r(m, k)$ can be represented by a form of the type $r(m, k) = a(k)\, m + b(k)$. The coefficients $a(k)$ and $b(k)$ have been evaluated for each of the k's in the range of linearity. Below is given the form which is perhaps of greatest immediate interest

$$r(m, 2) = 0{\cdot}141965\, m - 0{\cdot}137709 \quad . \quad . \quad . \quad (31)$$

When plotted, this function has a maximum deviation of 3·2 per cent from the exact curve for $r(m, 2)$. For example with $m = 4/3$, the above expression yields the value 0·051578. Rounded to 4 figures, this gives 0·0516 compared to 0·0513 for the exact reflectance, with a deviation of about 0·6 per cent.

In general a good rule of thumb for finding $r(m_2, k)$ if $r(m_1, k)$ is known is given by

$$r(m_2, k) = r(m_1, k) + 0{\cdot}142\,(m_2 - m_1),$$

$$\tfrac{1}{2} \leqq k < \infty, \quad 1{\cdot}2 \leqq m \leqq 1{\cdot}9. \quad . \quad . \quad . \quad . \quad (32)$$

The maximum errors of deviation in percentages over the range $1{\cdot}2 \leqq m \leqq 1{\cdot}9$ for each k is given by the following pairs of numbers (6·9 per cent, 0·5), (5·4 per cent, 1), (3·2 per cent, 2), (2·4 per cent, 2·5), (1·7 per cent, 3), (1·2 per cent, 3·5), (0·7 per cent, 4), (0·3 per cent, 4·5), (0·1 per cent, 5), (6·0 per cent, ∞).

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Annual rainfall probability maps of the British Isles

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Summary

Maps showing the percentage probability of receiving an annual rainfall of less, or more, than 20 in., 30 in., 40 in. and 50 in. are presented for the British Isles, and the conditions shown on these maps are discussed.

1. Introduction

Over the past half-century there have been innumerable studies of rainfall conditions in Britain but, as was stressed recently (Gregory 1957), no maps have yet been presented to show the probability of receiving given totals of annual rainfall. The aim of this paper is therefore to help fill this gap by constructing, and commenting on, four such probability maps for annual falls of 20 in., 30 in., 40 in. and 50 in. respectively.

The method employed in the preparation of these maps has already been described for East Africa (Glover, Robinson and Henderson 1954) and need not be repeated at length here. It involves, firstly, the calculation, for a considerable number of stations, of average and standard deviation values. From these, by the use of the appropriate formula and the requisite statistical tables, the percentage probability of an annual fall above or below the critical value may be computed. For a number of stations data, from which such calculations could be made, have in fact been in published form for over 30 years (Glasspoole 1925). However, although the assessment of probability for individual localities has been made from them, there has been no attempt to use them to depict the overall pattern.

The data analysed in the present paper are for the period 1901-1930, and consist of 230 station records extracted from the volumes of *British Rainfall*. These would have provided a picture of the broad differences within the country, but for the details of these differences, especially in upland areas, more data were required. For this purpose some 670 ' derived ' stations were obtained in the following way. The standard deviations for the 230 recording stations were converted into relative variability values by expressing them as a percentage of the average, and an isopleth map was prepared. This map was then superimposed upon a photographically enlarged copy of the 1901-1930 mean annual rainfall map from the *Climatological Atlas of the British Isles*. Each intersection of isopleths from these two maps was then used as a ' derived ' station, with its standard deviation readily obtained from the average and relative variability data.

The choice of the period 1901-1930 was partially conditioned by the availability of adequate data, and partially by the desire to produce maps which could be used in conjunction with those for this standard period available in the *Climatological Atlas of the British Isles*. The actual period of years is important in another way, however, for the 1901-1930 period included a greater proportion of wetter years than does the full period since reliable observations were started (Gregory 1956). This means that, in the maps which follow, the tendency is for the frequency of falls less than any given critical value to be somewhat underestimated, and the frequency of falls greater than any given critical value to be partially overstressed, in comparison with conditions over the full observational period.

2. The probability maps

In all these four maps a standard form of presentation has been employed. The probability values are expressed as percentages, and the key is phrased so as to indicate the probability of receiving both more than, and less than, the given value. Also, in all except the 20 in. map, a common grading and shading system has been used to facilitate direct comparison. On the 20 in. map, however, one category (a 10-30 per cent probability of receiving less than 20 in.) has been subdivided, as the range of values involved on the map is small.

It will be readily seen that most of the grades occur across only a very narrow belt of territory, with a rapid change from 10 per cent to 90 per cent values. This reflects the interaction between the relatively low variability of annual rainfall over much of Britain and the rapid change in rainfall totals between neighbouring localities as a result of relief contrasts. Conversely, the 0-10 per cent and 90-100 per cent grades cover relatively large areas in all cases, this being largely a reflection of the statistical method employed. Thus in the 0-10 per cent grade, for example, the chances of receiving the given amount vary from 1/10 to over 1/1,000, and for the bulk of the area the chances are nearer the latter than the former. It is only near to the 10 per cent line that chances not exceeding 1/100 are likely, so that much, though not all, of this 0-10 per cent grade can virtually, if not statistically, be grouped with the area of 0 per cent.

(*a*) *The 20 in. probability map*

For the period 1901-1930 no station employed received an average annual rainfall of less than 20 in. Many areas, however, had values sufficiently near it for some individual years within the period to have falls below this amount, whilst yet others suggested the possibility of such low falls some time during a longer period. The chance of a fall of below 20 in. at least one year in ten occurs over only limited areas (Fig. 1), the major expanse extending along the east coast between the estuaries of the Humber and the Thames, and inland to the lower Trent valley, the fringes of the east Midlands and into Oxfordshire. Within this area, some localities can expect such low falls rather less frequently (about an 8 per cent probability), whilst conversely along the Essex and Suffolk coasts, and around and south of the Wash, a probability of two or even three years in ten occurs.

Elsewhere over the country a probability of 10 per cent occurs locally, as around the lower Tees and over the north-eastern tip of Aberdeenshire, but probabilities soon fall to the order of 1/100 or less. Finally, as one moves into the wet highland zone, or into areas of only moderate rainfall combined with low variability as over much of Ireland, even the statistical possibility of a fall of less than 20 in. in a year ceases to exist.

(*b*) *The 30 in. probability map*

The chances of receiving less than 30 in. of rain in a year are much greater over considerable areas of Britain, and a far larger area is affected (Fig. 2). Over the east-coast lowlands, especially

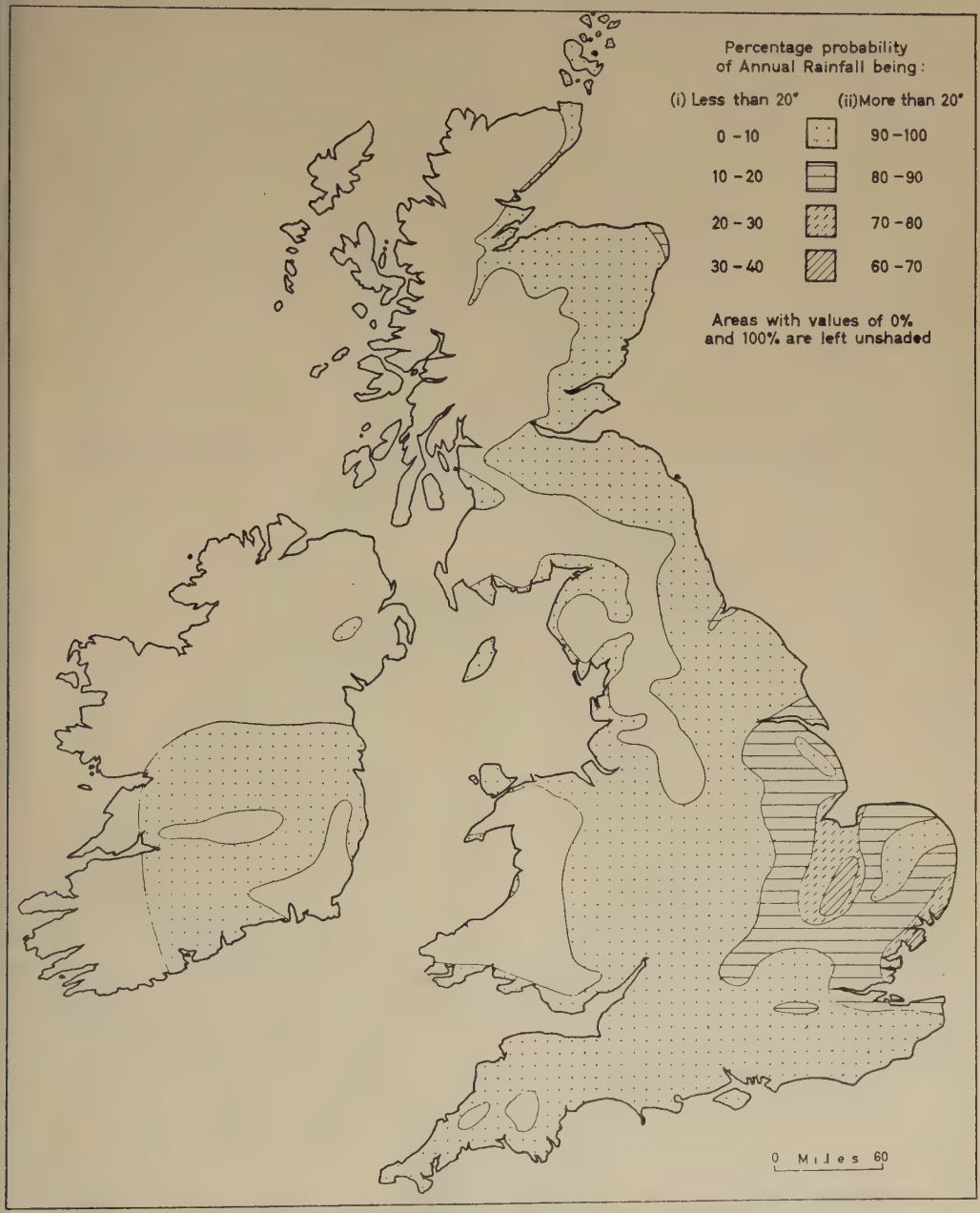

Figure 1. The percentage probability of receiving an annual fall of less, or more, than 20 in.
(Figs. 1-4 based on data from *British Rainfall* 1901-1930 and the *Climatological Atlas of the British Isles*).

the estuary areas, a fall of below 30 in. will occur in 90-100 per cent of the years – the Essex coastal belt, the valleys draining into the Wash and the Humber, the lower Tees and the lower Tweed, and part of coastal Nairnshire and Morayshire, all fall into this category.

In the northern half of Britain the chance of receiving less than 30 in. decreases rapidly towards the higher land. Probabilities of 10 per cent or less occur over the uplands, whilst on much of them rainfall exceeds 30 in. every year. This also holds true farther south over Wales and south-western England, but the transition from this extreme to the reverse around the Wash is much more gradual than it is in the north. Values of between 50 per cent and 90 per cent probability of less than 30 in. occur over the eastern half of southern England as far south as the Thames valley, with extensions of comparable conditions along the Avon and Severn valleys and over Wirral. Finally, the relative uniformity of 0-10 per cent values over Ireland is modified

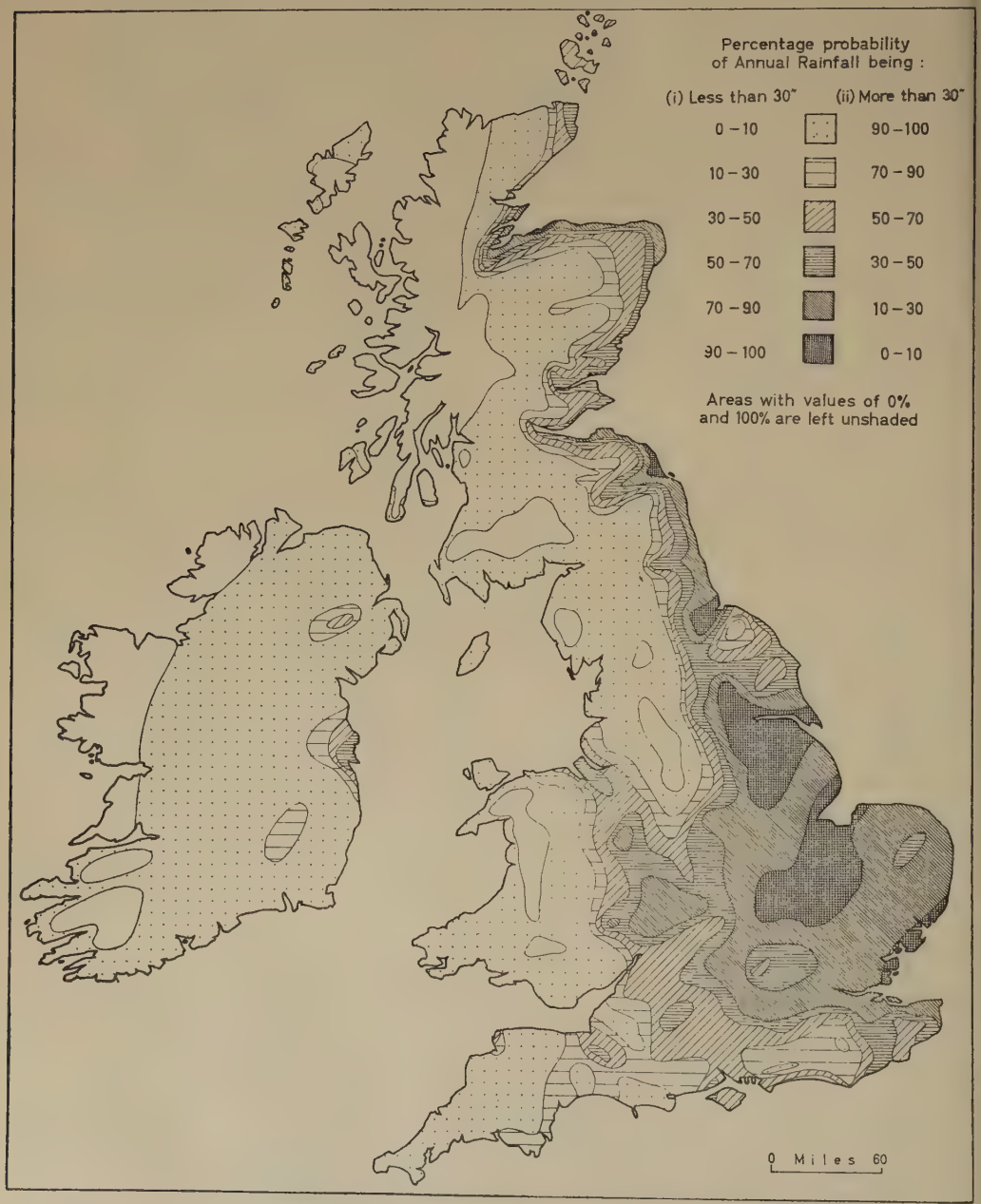

Figure 2. The percentage probability of receiving an annual fall of less, or more, than 30 in.

nly by the occasional area of more frequent low rainfall in the east, and by the absence of falls elow 30 in. in the far west.

In many ways the most significant point is that in no part of the country is a year's fall of ver 30 in. an impossibility, whilst conversely in only limited areas will a fall of less than this mount never occur. It should also be noted that around the fringes of the hill areas of Britain, here average annual rainfall may well be 35-40 in., a fall of less than 30 in. occurs between ne and three years in ten.

(c) *The 40 in. probability map*

When the probability of a fall of 40 in. is considered (Fig. 3), one finds not only areas where his total is never reached (the east coast lowlands again) but also areas where this total is always

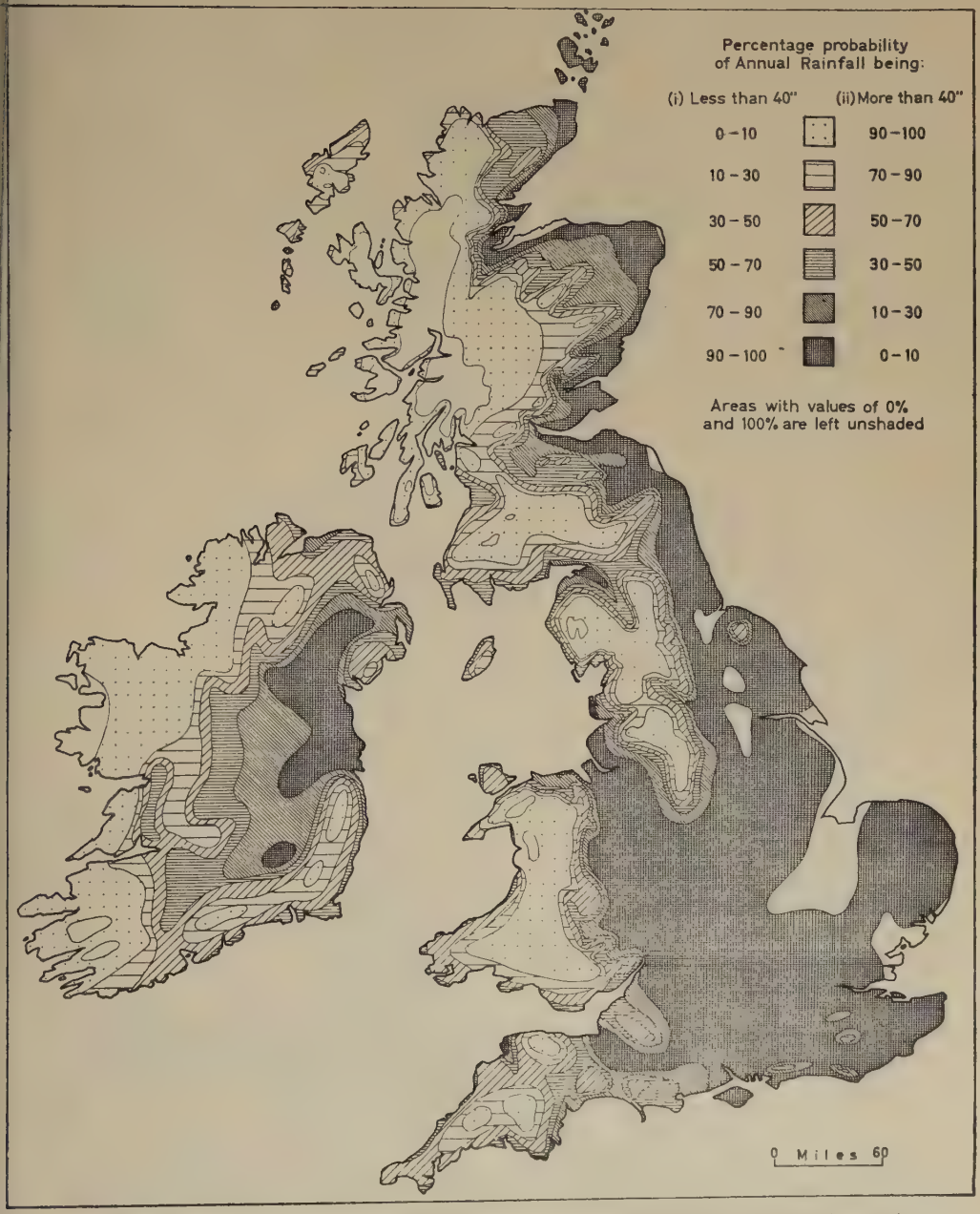

Figure. 3. The percentage probability of receiving an annual fall of less, or more, than 40 in.

exceeded (the highest parts of upland Britain and a few peninsulas in Ireland). Between these two extremes, the largest area experiences less than 40 in. at least nine years in ten. In England, Scotland and Wales there is then a rapid transition, over the lower slopes of the hills, to the upland expanses where a fall of over 40 in. occurs at least nine years in ten. In fact, these countries could be divided fairly equally into two broad zones, the one characterized by a 90 per cent or greater chance of receiving less than 40 in. of rain per annum, and the other by a 90 per cent or greater chance of receiving more than this amount – the transition zone between them is very narrow indeed. In Ireland, however, the generalization does not really hold true, for values change more gradually westwards and into the isolated hill areas.

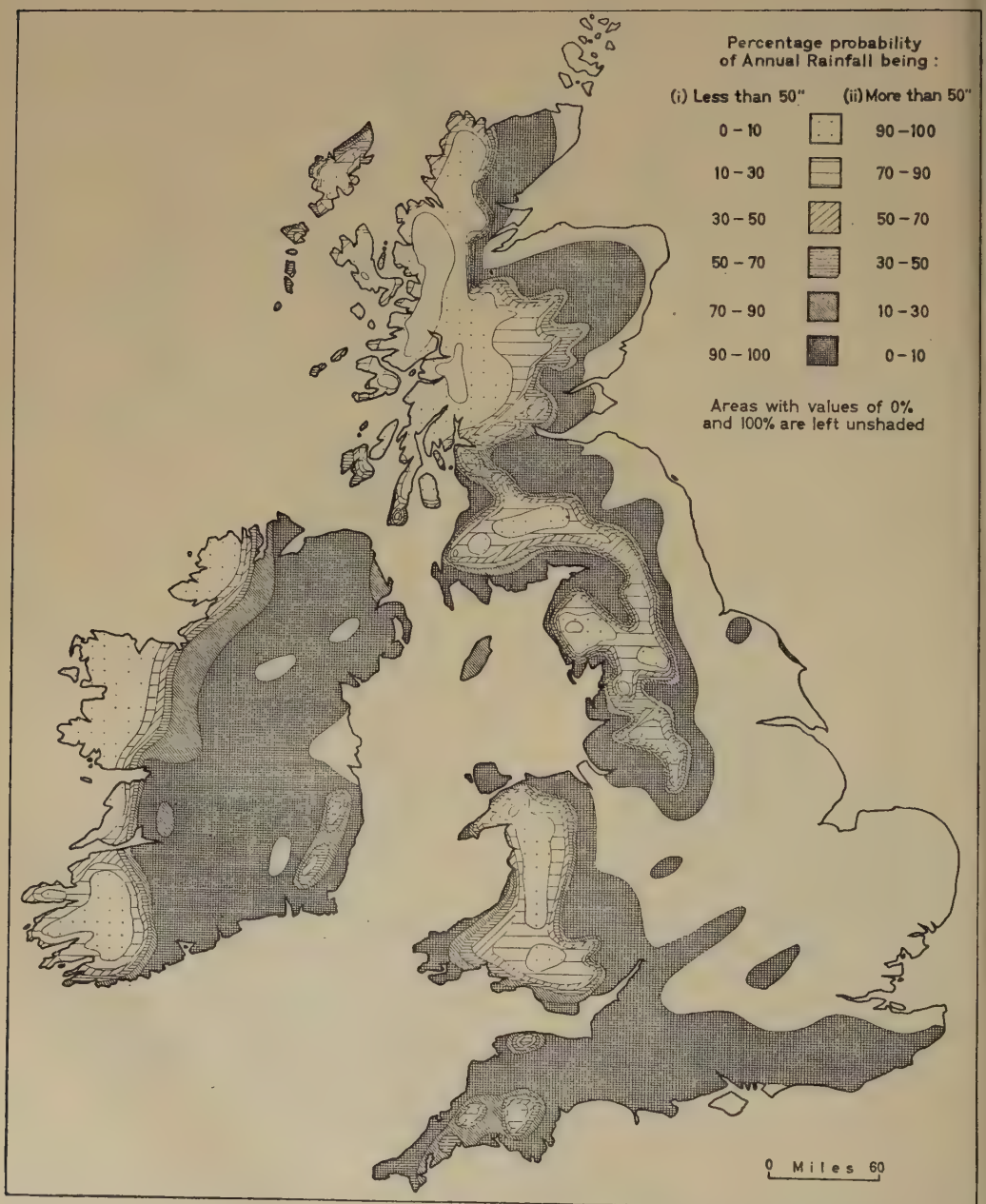

Figure 4. The percentage probability of receiving an annual fall of less, or more, than 50 in.

(*d*) *The 50 in. probability map*

This contrast based on the two 90 per cent isopleths is equally striking in relation to the probability of 50 in. (Fig. 4), but the most obvious difference is the large area over which there is clearly no chance of receiving the given total of rain. Also a chance of less than 1/10 of receiving this amount now covers the lower slopes and valleys of the hills, extends over most of the south-west Peninsula, and affects the bulk of Ireland. Thus the gradient between this set of conditions and a 90 per cent chance of receiving the given amount, has everywhere been intensified on the 50 in. map. These areas receiving over 50 in. at least nine years in ten coincide, to a remarkable degree, with the areas which always receive more than 30 in. (Fig. 2). Furthermore, the areas with less than 50 in. at least nine years in ten (Fig. 4) are, to a slightly lesser degree, comparable with those areas which are likely, at some time, to experience a fall of less than 20 in. (Fig. 1).

3. Conclusions

These four maps thus present the overall pattern of annual rainfall probability over the British Isles, within the limits imposed both by the data and by the methods used. The general relationship with the distribution of mean annual rainfall is only to be expected. In detail, however, the differences in the standard deviation, varying as it does between less than 10 per cent and more than 20 per cent of the mean value, introduce marked divergences from this mean annual rainfall pattern. Thus the peninsulas of western Ireland, with their low relative variability, display in the maps presented here a closer relationship with the wettest areas of upland Britain than a study of their average rainfall would suggest. Equally, in the upper and lower grades of the probability scale, areas of markedly different mean rainfall are grouped in the same percentage grade. In this way, patterns are created which are both simpler, and yet in many ways more significant, than that provided by the mean annual rainfall map.

References

Glasspoole, J.	1925	*British Rainfall*, **65**, p. 254.
Glover, J., Robinson, P. and Henderson, J. P.	1954	*Quart. J. R. Met. Soc.*, **80**, p. 602.
Gregory, S.	1956	*Geographical Journal*, **122**, p. 346.
	1957	*Advancement of Science*, **13**, p. 347.

551.552 (42)

The frequency of light winds in the English Channel

By G. SPURR

Central Electricity Authority, Leatherhead, Surrey

(Manuscript received 12 April 1957, in revised form 18 June 1957)

1. Introduction

Plans have been made to inter-connect the electricity systems of Britain and France by means of a cable across the English Channel. Once the cable has been installed the possibility of breakdowns must be considered and, in particular, it would be useful to know in advance the chances of carrying out repairs at all times of the year, bearing in mind that repair operations can only be carried out in relatively calm weather.

The information available was found to be somewhat sparse and a statistical analysis had to be carried out in order that the available data could be applied to the particular requirements of the problem. The results of this investigation are presented in this paper, as it is thought that they may be of interest in other applications.

2. Operational requirements

If a fault occurs in the cable, it is assumed that the following periods will be required to carry out a repair : One day for the location of the fault from the shore; three days for the location of the fault at sea; two days to repair the first joint; two days to repair the second joint : i.e. seven days for the total operation at sea.

In terms of limiting wind speeds the phases of the operation may be divided as in Table 1.

TABLE 1.

Beaufort	Speed (kt)	Phase of operation
0 – 2	0 – 6	All operations possible
$\leqslant 3$	$\leqslant 10$	Surface work possible, but not joint repair
$\geqslant 4$	$\geqslant 11$	No sea operation possible

From an analysis of wind frequencies, the most useful information is likely to be :

(a) the probability of two or more consecutive days with light winds in any one month (0-6 kt).

(b) the probability of at least seven days with winds less than or equal to 10 kt (not necessarily consecutive) in any one month.

3. Data

Data can seldom be found which are in a form directly applicable to a particular problem, and it has been found necessary to use statistical methods to obtain the required results from the information available.

3.1. Wind frequencies taken from the German Marine Atlas

Data abstracted from the *German Marine Atlas* (*1954*) are given in Table 2.

TABLE 2. Percentage of time the wind blows at the given force (Beaufort scale)

Dover Straits

Month	Calm	1-3	4-5	6-7	8-12
Jan.	2	38	34	18	8
Feb.	2	39	35	17	7
Mar.	5	43	32	15	5
Apr.	5	51	31	11	2
May	6	62	25	6	1
June	8	59	27	5	1
July	8	61	25	5	1
Aug.	6	56	30	7	1
Sept.	5	55	31	7	2
Oct.	3	43	35	14	5
Nov.	2	36	35	18	9
Dec.	1	32	37	22	8

The frequencies given in Table 2 have been expressed as the number of days per month the wind blows on average at the given force (Table 3). This involves the broad assumption that the wind force usually persists in one of the ranges given for at least one day. A comparison with four months of actual hourly reports of wind at Dover suggests that the results are not unreasonable.

4. Statistical analysis

From the data given in Table 3 cumulative frequencies were plotted for each month and from these graphs Table 4, giving the number of days the wind blows at less than or equal to 10 kt and less than or equal to 6 kt, has been compiled.

TABLE 3. NUMBER OF DAYS PER MONTH THE WIND BLOWS AT THE GIVEN BEAUFORT FORCE

Month	Calm	1-3	4-5	6-7	8-12
Jan.	½	12	10½	5½	2½
Feb.	½	11	10	4½	2
Mar.	1½	13	10	5	1½
Apr.	1½	15½	9½	3	½
May	2	19	8	1½	½
June	2	18	8	1½	½
July	2	19	8	1½	½
Aug.	2	17	9½	2	½
Sept.	1½	16½	9½	2	½
Oct.	1	13	11	4½	1½
Nov.	½	11	10½	5½	2½
Dec.	0	10	11½	7	2½

TABLE 4. NUMBER OF DAYS PER MONTH THE WIND BLOWS AT THE GIVEN SPEED

Month	$\leqslant$ 10 kt	$\leqslant$ 6 kt
Jan.	12½	7
Feb.	11½	6
Mar.	14½	9½
Apr.	17	11
May	21	14½
June	20	13
July	21	14
Aug.	19	13
Sept.	18	11½
Oct.	14	8½
Nov.	11½	6
Dec.	10	5

TABLE 5. EXPECTED NUMBER OF RUNS OF TWO OR MORE DAYS WITH WIND SPEEDS $\leqslant$ 6 kt

	Number of runs	
Month	E	Nearest whole number
Jan.	1·84	2
Feb.	1·51	2
Mar.	3.03	3
Apr.	3.83	4
May	5·41	5
June	4·79	5
July	5·20	5
Aug.	4·75	5
Sept.	4·08	4
Oct.	2.54	3
Nov.	1·44	1
Dec.	1·01	1

If the probability of any day having a light wind is p, and the probability of any (and in particular of the preceding) day having a strong wind is q, then the chance that the day begins a sequence of light-wind days is pq. The probability that the following day has a light wind is p, therefore the probability of a run of at least two days with light winds is qp^2. Hence in a month of N days, the expected number of sequences of two or more days with light winds is

$$E = Nqp^2$$

This result assumes that there is no tendency for persistence; but with weather phenomena there is a tendency for a condition to persist once started (Brooks and Carruthers 1953). The best estimate for the persistence factor is about 1·5 so that

$$E = 1{\cdot}5\ Nqp^2$$

Using this relationship with the data from column 3 of Table 4, the expected number of sequences of two or more days with light winds has been derived for each month (Table 5).

4.1. The probability of not less than 7 days with winds less than or equal to 10 kt in any month

The probability that any one day in a month will have a wind speed of less than or equal to 10 kt is obtained from column 2 of Table 4 by dividing the average number of days with wind less than or equal to 10 kt by the number of days in the month.

The probability that there should be not less than 7 such days in any month is given by :

$$P(k, N) = 1 - \sum_{m=0}^{k-1} \frac{N!}{m!\,(N-m)!}(p)^m (q)^{N-m}$$

where

$k = 7$ days

$N =$ Number of days in the month

$p =$ Probability of a day with wind $\leqslant 10$ kt

$q =$ Probability of a day with wind > 10 kt

By inspection of Table 4, the windy months appear to be November, December, January and February. The above formula has been applied to these months and Table 6 has been derived from the results.

TABLE 6. PROBABILITY OF NOT LESS THAN 7 DAYS IN ANY MONTH WITH WIND SPEED $\leqslant 10$ kt

Month	Percentage chance
Dec.	91
Nov.	97
Feb.	97
Jan.	99
All other months	> 99

5. CONCLUSIONS

The limited amount of data relating to winds in the Dover Straits have been analysed, with a view to assisting in the estimation of possible maintenance of the Cross-Channel cable at any time of year. For most of the year there appear to be sufficient periods of light winds to allow work to be carried out. During the period from November to February, however, there is some risk that the necessary number of suitable days will not occur in any one month. During the months of November and December, in particular, it appears likely that the favourable conditions will not arise, there being an expectancy of only one run of two or more days of light winds, whereas two runs are required.

It has been assumed that a repair could be carried out in seven days. Experience may show that a longer or shorter period will be required. In this case different conclusions will be drawn, but the methods and data contained in this report will still apply.

ACKNOWLEDGMENT

This paper is published by permission of the Director of the Research Laboratories, Central Electricity Authority.

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Deutscher Wetterdienst 1954 *Klimatologie der Nordwesteuropäischen Gewässer*, Seewetteramt, Hamburg.

Correspondence

551.501.7 : 551.557.36 : 551.547.3 : 519.2 : 551.515

The relation between standard deviation of contour heights and standard vector deviation of wind

By P. D. Thompson

I read with great interest A. F. Jenkinson's recent paper on the relation between the standard eviations of wind and height (*Q.J.*, **82**, p. 198), and would like to point out certain interpretations f his results that were either overlooked or partially obscured in his discussion.

The main point is that the quantity k – an empirically determined factor of proportionality etween the standard deviations of wind and height – is, in fact, a statistical measure of the *orizontal scale* of deviations from average conditions. To show this, let us consider a characteristic wave number ' m, defined as

$$m^2 = -\frac{\overline{z\,\nabla^2 z}}{\overline{z^2}}$$

here z is the deviation from the climatological mean, ∇^2 is the horizontal Laplacian operator, nd a ' bar ' placed above a quantity denotes its time average. The quantity m has the units of a eciprocal length, and is relatively independent of the amplitude of deviations from the mean. t takes on large values when the horizontal variations of z are large – i.e., when for deviations f given amplitude, the horizontal scale is small – and vice versa. For example, if z were a inusoidal function of a single coordinate, m would reduce to the true wave number. Now, y definition,

$$m^2 = -\lim_{\epsilon \to 0} \frac{1}{\epsilon^2} [\overline{z(x,y)\,z(x+\epsilon,y)} + \overline{z(x,y)\,z(x-\epsilon,y)} + \overline{z(x,y)\,z(x,y+\epsilon)} + \overline{z(x,y)\,z(x,y-\epsilon)} - 4\,\overline{z^2(x,y)}]/\overline{z^2(x,y)}$$

n which x and y are horizontal cartesian coordinates. Thus, assuming (with Jenkinson) that the tatistics of z are isotropic,

$$m^2 = 4 \lim_{\epsilon \to 0} \frac{1}{\epsilon^2}(1 - r)$$

where r is the correlation between z values at points ϵ apart. The limit above generally exists f $d^2 r/d\epsilon^2$ exists at $\epsilon = 0$.

It will be noted that Jenkinson's k is proportional to our m by a constant factor of $g/2\omega$. Thus, Tables 3 and 4 of Jenkinson's article do not testify merely to the quasi-constancy of an mpirically determined parameter, but provide valuable (and hard won) information about the easonal, geographical, and vertical distributions of the typical scale of fluctuations in atmospheric low patterns.

The most striking single feature of Jenkinson's results is, of course, the remarkable constancy f k throughout the troposphere, for different regions and varying seasons. Averaged over all tations and seasons, the values of k at elevations ranging from 850 to 300 mb deviate by less han 3 per cent from a grand average of 0·0635 kt/ft, a figure that corresponds to a ' wavelength ' f about 2,400 mi. This fact is rather surprising, for it implies that the characteristic horizontal limension of *transient* pressure systems at the surface is about the same as that of pressure fluctua- ions at all levels up to the tropopause. It suggests, further, that the migratory systems may ossess more vertical coherency than is usually attributed to them. It would be interesting to ind out whether many of the small-scale details of surface pressure patterns are really significant eatures of the *migratory* systems, or whether they are due to the superposition of *semi-permanent* rregularities in the surface pressure pattern – as, for instance, the variations in the climatological listribution of ' sea-level ' pressure over the western U.S. in winter.

Jenkinson has also presented fairly convincing evidence that k decreases markedly above the ropopause, from which we might infer an approximate doubling in the scale of height fluctuations s one proceeds upwards to 100 mb. By and large, this result is in accord with extant theories of

baroclinic instability, all of which state that the wavelength of maximum instability (for disturbances of great vertical extent) increases as the static stability is increased. An alternative explanation (which I find difficult to justify quantitatively, and do not entertain wholly) is that an analyst's reconstruction of a contour pattern from data at a finite number of points becomes noticeably smoother as he is deprived of more and more data. If the former is the case, then the abrupt change of static stability across the tropopause and the attendant change in horizontal scale (wavelength of maximum instability) might be expected to result in a fundamental lack of coupling between the flow patterns above and below the tropopause. It is, of course, a matter of synoptic experience that the height patterns in mid-troposphere and mid-stratosphere are not very closely connected.

Laying speculation aside, I wish to draw attention to Mr. Jenkinson's valuable and very timely data on the characteristic scale of fluctuations in atmospheric flow patterns – a quantity that appears more and more frequently in studies of the statistical behaviour of hydrodynamical systems.

Development Section,
Joint Numerical Weather Prediction Unit,
U.S. Weather Bureau,
Washington.
6 March 1957.

Reply

By A. F. Jenkinson

In a preliminary draft of the paper I included a brief discussion showing how k was related to the scale of pressure systems, but this was omitted on advice from referees*.

Col. Thompson's discussion in terms of wave number is however more elegant and illuminating than was mine, and I thank him very much for his interesting comments.

From my own experience I would say that migratory pressure systems do possess considerable vertical coherency. We can see this if we examine the chief variations of the climatological mean picture. These variations should however not be regarded simply as pressure variations, but as changes in atmospheric motion which result dually in the establishment of certain pressure types and in the confluence of particular air masses.

In the *Handbook of Weather in the Mediterranean* (not yet published) I have considered seasonal weather types as arising dually from the movement of one or more of the quasi-permanent cells of high and low pressure and the confluence of particular air masses. When we thus examine the weather type we clearly see vertical coherency to the limit of convective effectiveness of air masses, viz., the tropopause. We note further that the daily weather maps for a considerable area can be assigned with relative ease to seasonal types.

I thus feel that the results of my paper on standard deviation of wind and contour height agree well with the results of recent synoptic research.

I also wish to point out that although vertical coherency of migratory systems is to be interpreted as being chiefly linked with major weather types, yet many so-called ' small-scale ' or ' local ' details of surface-pressure patterns are in reality bound up with a major weather type in a fundamental relation. I have illustrated this amply in the *Handbook of Weather in the Mediterranean*, and as examples of ' local ' weather which have an important relation with a major weather type I may mention (a) summer-morning low stratus in the Egyptian Delta, (b) summer sandstorms in the Euphrates Valley, (c) intensification and weakening of the 'etesians' of the Aegean, (d) summer fog in the Straits of Gibraltar. These phenomena can be related to wind fields near the tropopause. Other examples from winter weather in the Mediterranean, e.g., the mistral and bora, are well known.

The Waite Institute,
Adelaide, S. Australia.
3 June 1957.

* Not the Society's referees (Editor).

551.574.1 : 551.578 11

A theoretical estimate of the collection efficiencies of small droplets

By T. PEARCEY and G. W. HILL

(Read 20 March 1957. Presented by Mr. J. S. Turner. See this volume, p. 77)

Dr. B. J. Mason : Perhaps the possible significance of these calculations for cloud physics can best be raised by pointing out how they differ from the results of previous work. They appear to raise two main issues.

First, they suggest that for two droplets of comparable size and having values of $R > 1$, the collision cross-section E may be considerably greater than the corresponding geometrical cross-section because the larger droplet becomes sucked into the wake of the smaller. Thus the collision efficiency for a pair of drops having radii 31 μ and 30 μ is calculated to be 9, the corresponding values for droplets of 31 μ and 20 μ being unity, and for 31 μ and 10 μ droplets only 0·16. At first sight, this might suggest that the coalescence mechanisms should proceed quite rapidly in a cloud having a narrow drop-size distribution but, in fact, the high values of E for droplets of nearly equal size are largely offset by the difference in their fall velocities being very small.

These calculations of E may, of course, be challenged on the grounds that the Oseen solution is not reliable in the neighbourhood of the droplet surface, and that it is obviously not permissible to assume that the flow pattern produced by the approach of two nearly equal spheres can be represented by a linear superposition of the individual flow patterns, particularly when the spheres are close together. It is difficult to judge to what extent the calculated values of E will be influenced by these assumptions without solving the two-body problem rigorously. Would the authors expect their assumptions to lead to an overestimate or an underestimate of E for two droplets of comparable size? It may be of interest to quote the results of some recent model experiments of Schotland and Kaplin (1956), in which the motion of two nearly equal-sized cloud droplets, with $R < 1$, were simulated by ball bearings falling in a viscous sugar solution. Their relative trajectories in three dimensions were constructed from photographs taken in two mutually perpendicular planes. For a simulated cloud droplet of 10 μ radius colliding with one of 8 μ radius, Schotland and Kaplin found a collision efficiency of 0·16, while Pearcey and Hill's calculated value is 0·04. For a droplet pair with radii 8·5 μ and 5 μ, the experimental value of E was 0·08; Pearcey and Hill give zero.

The second point of importance is that for a drop approaching a much smaller one, the values of E calculated by Pearcey and Hill are considerably smaller than we have hitherto assumed on the basis of Langmuir's computations. If one corrects Langmuir's results for the finite size and fall velocity of the smaller drop, the disagreement with Pearcey and Hill becomes even worse. Comparative values are shown in the following table :

a_1	a_2	E_p	E_v	E_L	E'	E''
200 μ	7·6 μ	0·92	0·50	0·67	0·25	0·98
73 μ	5·8 μ	0·84	0·26	0·30	0·04	0·20

where a_1, a_2 are the droplet radii, E_p, E_v are respectively the corrected Langmuir values for potential and viscous flow, E_L the value obtained by using Langmuir's interpolation formula, E' is Pearcey and Hill's value, and E'' represents the experimental values given recently by Kinzer and Cobb (1956). There is obviously a wide measure of disagreement, Pearcey and Hill's values being markedly lower than the others.

In the past it has been customary to assume that conditions near the front surface of a drop of large Reynolds number approximate to those of potential flow, when collision efficiencies for very much smaller droplets become of order unity. The Oseen solution, adopted by Pearcey and Hill, which predicts the formation of a boundary layer at the front surfaces of drops for which $R > 10$, will obviously give considerably lower values of E since the streamlines in the neighbourhood of the sphere are deflected outwards by an amount comparable with the thickness δ of the boundary layer, where $\delta \simeq 2a_1/R_1^{\frac{1}{2}}$. Thus droplets with $a_2 < \delta$ will suffer an appreciable deflection away from the front surface of the larger drop.

In assessing the implications of this for cloud physics, one may ask whether the Oseen solution, which is undoubtedly an improvement on the Stokes solution at distances remote from the sphere and for values of $R < 3$, may not be seriously in error near the surface of the sphere and for large Reynolds numbers. Since it does not seem possible to assess the errors which will arise from such simplified mathematical treatments of the problem without undertaking a rigorous numerical solution of the full hydrodynamical equations, it seems highly desirable to settle the matter by experiment, although the difficulties of measuring the collision efficiencies of small droplets of nearly equal size are rather formidable. Meanwhile we may draw what little comfort we can from the experiments of Gunn and Hitschfeld (1951) which indicated the collection efficiency of millimetre-size drops for droplets of 5-50 μ radius to be close to Langmuir's values for potential flow and from the fact that the same is true for the collection of lycopodium spores ($r \simeq 15\,\mu$) by sticky cylinders (Gregory 1951). Personally, I am inclined to think that for large values of R and small values of a_2/a_1, the potential flow solution is more realistic for conditions at the front surface of the large drop than is the Oseen solution, although, of course, it cannot represent what happens at the rear.

Dr. T. Pearcey and Mr. G. W. Hill (*in reply*) : In general we agree with Dr. Mason's remarks and criticisms of our paper which was specifically restricted to hydrodynamical factors likely to affect the rate of collision between cloud droplets. Little was said about their influence on the actual coalescence rate within a cloud.

In regard to Dr. Mason's first remark, the coalescence rate between droplets of nearly equal size is partly offset by their slow rate of approach, but their contribution is still augmented by the wake effect in the absence of microturbulence and surface effects which may reduce the effective collection efficiency.

Although the Oseen approximation to the flow pattern shows flow characteristic of a boundary layer near the forward surface of a droplet, it is just in this region where we expect the approximation to diverge most from reality. It is possible that it does not approach the potential flow rapidly enough with increasing R, and would thus partly account for the small values of E found by us compared with Langmuir and indicated by experiment (see, for instance, Stewartson, *Phil. Mag.*, 1956).

Linear superposition of the fields of flow of two droplets is expected to be satisfactory in most cases, when far apart, when the strength of one field far exceeds that of the other, as in approach down a wake, and when relative velocities are large. Failure of linear superposition may be expected to affect droplets of very nearly equal size, especially small sizes, while neighbouring pairs of droplets lie close together for relatively long periods. In the case of larger droplets the fields are of shorter range and the droplets of larger mass and less likely to be affected. At close approach further effects must be considered, as for instance, the ' rolling ' which may arise from the torque experienced by a droplet in a field of flow possessing shear. How these factors affect trajectories is not yet known.

We agree with Dr. Mason that the small values of E computed by us for large size ratios may be due partly to inadequacy of Oseen's approximation, of linear superposition and the omission of effects such as ' shear rolling.' In view of the mathematical difficulties in allowing for these more complex effects, an attempt at experimental simulation seems to be called for. Experiments of the Kinzer and Cobb type may provide answers to some questions about hydrodynamic effects, but we believe that true similitude must be maintained with liquid droplets.

Proceedings at meetings of the Society

Special General Meeting, 19 June 1957

A Special General Meeting of Fellows was held to elect the Officers and Council for the year 957-58.

Fellows had been circulated on 10 April with the list of nominations by Council for Officers nd Council for the year 1957-58. No other nominations having since been received, the persons o nominated were declared to be elected unopposed in accordance with By-law 51.

The Officers and Council for 1957-58 were therefore elected as follows :

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VICE-PRESIDENTS

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CHARLES HENRY BRIAN PRIESTLEY, M.A., Sc.D., F.Inst.P.
REGINALD COCKCROFT SUTCLIFFE, O.B.E., Ph.D., F.R.S.
RICHARD GEORGE VERYARD, B.Sc.

VICE-PRESIDENT FOR CANADA

ROBERT CAMPBELL GRAHAM

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GEORGE DAVID ROBINSON, Ph.D., F.Inst.P. (Editor)
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JOHN STANLEY SAWYER, M.A.
CHARLES EDWARD WALLINGTON, M.Sc.

Ordinary Meeting, 19 June 1957

The following candidates were declared to have been elected that day by the Council :

FELLOWS

CASSWELL, Stanley Alan, Officers' Mess, R.A.F. Manby, Louth, Lincs.
FORSDYKE, Donald, B.Sc., A.R.C.S., 34 Glover Road, Pinner, Middx. (*transfer from Student Membership*).
JOHNSON, Russell Farrington, 6 Brook Street, Buxton, Norwich, Norfolk (*transfer from Student Membership*).
KRISHNA RAO, Potarazu, B.Sc., M.Sc., M.S., Department of Meteorology, New York University, New York 53, U.S.A.
THOMSON, D. A., Auchmuty Schoolhouse, Glenrothes, Fife, Scotland (*reinstatement; resigned 1949*).
WALKER, Peter Henry, Instr. Lt., R.N., " Windover," 38 Marsh Lane, Crosspool, Sheffield 10, Yorks. (*transfer from Student Membership*).

FOREIGN MEMBERS

BROWN, Rodger A., 27 Eckerson Avenue, Akron, New York, U.S.A.
FLEISHER, Aaron, Department of Meteorology, Massachusetts Institute of Technology, Cambridge 39, Massachusetts, U.S.A.

STUDENT MEMBER

ABANDAH, Ali, Department of Meteorology, Imperial College, Exhibition Road, London, S.W.7.

CANADIAN BRANCH

ASSOCIATE MEMBER

WILSON, Cynthia Violet, Department of Geography, McGill University, Montreal, P.Q., Canada.

The following papers were read and discussed :

1. (a) ' Airflow over an isolated hill ' by R. S. Scorer (*Q.J.*, **82**, p. 75) ;
   (b) ' Waves in the lee of an isolated hill ' by R. S. Scorer and M. Wilkinson (*Q.J.*, **82**, p. 419).
2. ' Airflow over mountains : the lee-wave amplitude ' by G. A. Corby and C. E. Wallington (*Q.J.*, **82**, p. 266).
3. ' Water-vapour flux and flux-divergence over southern England : summer 1954 ' by J. W. Hutchings (*Q.J.*, **83**, p. 30); presented by Mr. J. K. Bannon.

Summer Meeting, Glasgow, 3-5 July 1957

The fourth two-day Summer Meeting of the Society was held in the Department of Natural Philosophy, University of Glasgow, by kind permission of the University Court.

The meeting was formally opened by Mr. J. R. Atkinson, Vice-President for Scotland, and welcomed by Professor P. I. Dee, F.R.S., head of the Department of Natural Philosophy. Dr. T. W. Wormell, Vice-President, expressed the Society's thanks for the hospitality afforded.

The meeting consisted of two scientific symposia.

Professor P. I. Dee was in the Chair at the first symposium on ' Cloud Physics.' He welcomed Professor C. T. R. Wilson, F.R.S., Honorary Member, who was present. Papers were read by Dr. B. J. Mason on ' Recent developments in laboratory studies of cloud physics,' Mr. R. J. Murgatroyd on ' The microphysical structure of clouds sampled from an aircraft,' and Dr. T. W. Wormell on ' Cloud electrification processes and some of their effects.'

Dr. H. L. Penman, Secretary, took the Chair at the second morning's session on ' Problems of hill climate.' The chief contributors were Mr. R. Cranna on ' Outline of problem ' ; Mr. R. W. Gloyne on (i) ' Classification of dissected country ' and (ii) ' Variation of temperature and length of growing season with height ' ; Dr. D. H. N. Spence on ' Climate and vegetation ' ; Lt.-Col. P. D. Baird on ' Weather recording on Ben Macdhui ' and Mr. C. A. Halstead on ' Incidence and intensity of frozen soil in remote areas.'

During the tour of the Department of Natural Philosophy on the Thursday evening, the staff and research students explained the equipment and techniques. Professor Gordon Manley, Vice-President, expressed the appreciation of the Society.

Other events included a visit to O.W.S. *Weather Watcher* at Greenock on Thursday afternoon, and to the hydro-electric scheme at Loch Sloy on the Friday afternoon. Dr. H. L. Penman and Mr. R. G. Veryard, Vice-President, expressed the Society's thanks on these occasions.

The Council record their sincere gratitude to Mr. J. R. Atkinson and his colleagues on the Scottish Centre Committee who were responsible for the excellent local arrangements.

Summer Visit, 13 July 1957

The Summer visit of the Society was held on 13 July 1957 at the Atomic Energy Research Establishment, Harwell, by kind invitation of Sir John Cockcroft, O.M., F.R.S.

The visit had to be restricted to Members of the Society, as the Atomic Energy Research Establishment could not take a party exceeding thirty. Owing to the considerable demand a ballot was held to fill the available places.

After an introductory talk illustrated by a film, the party visited the isotopes and radio-chemical laboratories, and the B.E.P.O. nuclear pile.

At the close of the meeting a vote of thanks was moved by Professor Gordon Manley, Vice-President, who expressed appreciation of the excellent arrangements made, and the services of the guides.

Reviews

Dynamic meteorology and weather forecasting. By C. L. Godske, T. Bergeron, J. Bjerknes and R. C. Bundgaard. American Met. Soc., Boston and Carnegie Inst., Washington, 1957. Pp. xvi, 800; numerous Figs.; $15.00.

This book represents the continuation of research by the famous Scandinavian team of meteorologists which has been supported by the Carnegie Institution. It is a direct continuation of the work published by V. Bjerknes and his collaborators in *Dynamic meteorology and Hydrography* (Washington 1910 and 1911) and *Hydrodynamique physique* (Paris 1934). These two books have had a great influence on the development of meteorology, and the same theoretical and logical approach to meteorological problems is evident throughout the present volume.

Although *Dynamic meteorology and weather forecasting* has only appeared this year (1957), the manuscript was completed in 1948. Revisions and additions were made until 1951 when the manuscript was passed to the publishers, but C. L. Godske who was responsible for most of the dynamical meteorology had no opportunity to amend his contribution after 1948.

That such a delay in printing should have occurred at a time when theoretical meteorology was developing rapidly can only be regretted. Turning the pages emphasizes how rapidly some branches of meteorology have developed since 1951. For example we find only a short and pessimistic reference to numerical methods of forecasting, and a treatment of precipitation theory which takes little account of the coalescence process of droplet growth.

This is not to say that the book is wholly outmoded, but its main value will be as a reference work regarding the state of the science at the end of World War II. It is well set out for reference purposes, although it is not always easy to find the meaning of symbols without turning back many pages. Research meteorologists will also find in the book a fund of ideas and theories which, although not in the stream of current research, may nevertheless be a valuable stimulus for further work. The book, however, cannot be recommended to the newcomer to the science – it is too vast and needs critical reading in the light of more recent work.

The book contains 18 chapters in 5 parts. The parts are briefly : Part I Thermodynamics and statics of the atmosphere – a thorough and classical treatment of the subject; Part II Kinematics of the atmosphere – an extension of the authors' previous works in a branch of meteorology rarely studied adequately; Part III Hydrodynamics of the atmosphere – many meteorologists may find the treatment novel and stimulating – it follows *Hydrodynamique Physique* but differs in approach from most modern works; Part IV Climatological and synoptic models – an exposition by one of the founders of synoptic analysis of what analysis should be; Part V Weather forecasting – a disappointing section largely because it is difficult to see in it a practical scheme of forecasting; it combines a too elaborate technique of upper-air analysis and prognosis with a collection of forecasting rules which form no coherent scheme.

Clearly this is a book which should be available to all meteorologists – it represents the culmination of an epoch in scientific meteorology – but appearing at the present time it is likely to be little read.

J.S.S.

Scientific uses of earth satellites. Ed. J. A. Van Allen. Chapman and Hall, 1956. Pp. x, 316. 63s.

Here is indeed a sign of the times. Only a few years ago the apostle of the artifical satellite was mostly regarded as a crank or a visionary – depending largely on how polite one's nature – and his subject, astronautics, a matter of science fiction, strip-cartoon type for the most part. Yet perhaps by the time this review appears earth satellites will have been successfully launched and orbited, with observations made on and by them providing a not insignificant contribution to the corpus of the International Geophysical Year. Meteorologists must sometimes feel they already have a surfeit of observations and that, operational requirements apart, the observers should be given a long holiday and the analysts and synthesists left to work on what they already have without distraction of new material. Man, however, is by nature and can afford to be prodigal in his science and by being so throws up vast new conceits of which the earth satellite is not least in modern geophysics. The meteorologist, to be alive, must make himself aware of the possibilities of the new tool and give best thought to its exploitation for his science.

The book now under notice describes how United States geophysicists, in conference in January 1956, were addressing themselves to the possibilities of observations using earth satellites. There are 33 papers, mainly from physicists already experienced in upper-atmosphere research with rockets, and each paper is devoted to specific proposals for the scientific use of an artifical satellite. As the editor says in his preface, the evanescent nature of descriptions of proposed experiments is obvious, but much of the analytical work is of durable value and it seemed proper to publish the best early thinking on the subject. The reviewer, seeing the product, agrees.

There is no classification of the contributions to the conference but they mainly deal with the following : general problems of satellite instrumentation, tracking and the geophysical inferences – the form of earth and the density of air (from drag), spectra of radiation from sun and space, solar and cosmic ray particle analysis, meteor bombardment and interplanetary dust, geomagnetism and the ionosphere, and observations of cloud pattern, earth albedo and terrestrial radiation. The reviewer is not competent to pass all these papers under review – he has only skimmed the majority – and it would be inappropriate anyway to attempt this. The point of the book, for those not immediately engaged in rocket or satellite research, is in the stimulus and broadening of horizon that is offered, and he would be a dull soul who failed to profit thus from an acquaintance with these pages.

The conventional meteorologist, taken up with problems of weather and its evolution, is promised quite a time – in due course. A satellite at 2,000 miles altitude could at any instant survey an area of the order of the United States, and television of the cloud cover would thus make manifest the birth, evolution and decay of weather systems. The prospect is indeed intriguing. Meanwhile, satellites in closer orbits may be measuring the albedo of much smaller zones of the earth.

The mention of the evolution of weather leads to the geophysical *raison d'être* of a satellite as opposed to a rocket. The satellite needs a rocket to lift and eject it into its orbit so that the satellite has no merit over the rocket as far as exploration in the vertical is concerned. The satellite, however, provides the opportunity for the measurement of variations of a property with latitude and longitude and in time. Such variations are at the heart of most geophysical phenomena and so the satellite should be especially exploited to that end. But the instrumental problems will evidently be very serious in some cases, for horizontal and temporal gradients of geophysical quantities are often orders of magnitude smaller than their vertical gradients.

Here then is a book to browse over for anyone interested in atmospheric exploration, and much closely argued detail to be examined by anyone intimately concerned with the problems of measurement at high levels. The challenge to the instrument designer is indeed most exacting but one may be sure that it is the beginning of a new geophysical era which this book appropriately heralds.

P.A.S.

The Airglow and the Aurorae. Suppt. to Vol. V, *J. Atmos. Terr. Phys.* Eds. E. B. Armstrong and A. Dalgarno. London and New York (Pergamon Press), 1956. Pp. ix, 420; Figs.; Tables. £7 10s.

The Airglow and the Aurora are important fields in the study of the chemical and physical properties of the upper atmosphere in the programme of the International Geophysical Year and the object of an International Symposium on these subjects held in the Queen's University, Belfast, in Sept. 1955 was to present at this convenient stage a coherent survey of the most recent work on the observational, experimental and theoretical aspects of these phenomena. The 53 papers reproduced are classified into 8 sections, an introductory review by Prof. S. Chapman, observations of the airglow, observations of aurorae (a comparatively short section), theoretical considerations on the airglow, theoretical considerations on aurorae, laboratory studies, quantal calculations and instrumentation. Most of the papers are specialized and considerable knowledge of the subject is required before they can be appreciated. It is not intended as a text-book and some aspects are not mentioned, nor, in common with all similar publications, is it by now completely up to date. It is, however, an extremely desirable addition to the literature of these subjects and a landmark of the state of progress in late 1955.

In the observation of the airglow, the main task up to a few years ago was the identification of the substances responsible for the different emissions. This has been largely accomplished although further work is needed at infra-red wavelengths and the emphasis is now on the determination of the height of these emissions. Several papers indicate the considerable disagreements which now exist. Diurnal and seasonal variations are also receiving considerable attention, there are more accurate measurements of intensity available and promising efforts to determine the temperatures at emission levels are also described. The observational studies in the aurora include a review of radio investigations as well as several measurements in visible and infra-red bands. The central problem in understanding these phenomena is, however, the nature of the excitation mechanisms and about half the papers are concerned with theoretical interpretation and laboratory studies. Both of these are difficult and the results obtained up to date have only allowed limited conclusions to be drawn. Finally the section on instrumentation contains several interesting devices and new applications of which perhaps the most welcome to meteorologists is the possibility of measuring emission temperatures by use of interferometers. In this and other sections the use of high-altitude rockets as a powerful research aid is indicated – both for emission height measurement and the production of artifical sodium clouds around 100 km and it is to be hoped that further important results will be obtained with this tool in the near future. The production is excellent.

R.J.M.

Solar eclipses and the ionosphere. Suppt. to Vol. VI, *J. Atmos. Terr. Phys.* Eds. W. J. G. Beynon and G. M. Brown. London and New York (Pergamon Press), 1956. Pp. x, 330; Figs.; Tables. £7.

It was the study of changes in the ionosphere during solar eclipses in the 1930's that revealed conclusively that the ionizing agent in the E and F1 layers was electromagnetic rather than corpuscular radiation. Produced in the lavish style characteristic of its publishers and containing a series of papers read at a Symposium held in London in August 1955, this volume provides an excellent survey of the opportunities offered by the occurrence of a solar eclipse to study the physics of the upper atmosphere and the sun, and of the results of such investigations.

The papers are admirably arranged in eight sections which are preceded by a valuable introductory survey and summed up in a concluding short outline of the principal results, both by J. A. Ratcliffe. Summaries of discussions are given at the end of four sections.

The main section which consists of accounts of recent ionospheric eclipse investigations is followed by series of papers on the solar radiations that produce the ionosphere and on theories of the formation of the layers. D. R. Bates summarizes current views regarding the processes responsible for the formation of the

layers – D is probably caused by photoionization of nitric oxide by the Ly α line; E mainly by X-rays; and both F1 and F2 partly by X-rays and partly by the photoionization of atomic oxygen. There follow sections dealing with determinations of recombination coefficients from eclipse observations, with recent work on the influence of eclipses on the geomagnetic field (introduced by S. Chapman), and on solar radio noise. Meteorologists will be specially interested in a tantalizingly short abstract of a paper on ' Chemical tests and the study of the upper atmosphere ' by G. Piccardi which appears in the concluding section of miscellaneous papers.

A comprehensive bibliography is provided at the end of the volume.

J.P.

Antarctica in the International Geophysical Year : Geophysical Monograph No. 1 of the American Geophysical Union. Baltimore (Waverley Press), 1956. Pp. v; 133; Figs. and Tables. $6.00.

In April 1956 the American Geophysical Union sponsored a symposium on Antarctica to stimulate interest in I.G.Y. activities there: the main papers read at the meetings are published in this monograph with the object of " bringing together available knowledge of the continent for background purposes in examining the major scientific problems. As a collection of available scientific knowledge about Antarctica the book is understandably (even necessarily) far from complete, and in some at least of the fields of inquiry the problems are skimpily treated: it is not a handbook for the expeditions that will spend the I.G.Y. on the continent. But within its scope the monograph presents a useful summary of the more important themes of geophysical and allied interest in this comparatively unknown region of the globe, where 86 per cent of the world's glacier ice is locked up, and where 11 nations will operate some 35 stations from now until the spring of 1959.

In two introductory chapters J. Kaplan, (Chairman of the U.S. National Committee for the I.G.Y.), and H. Wexler impressively describe the United States contribution to the I.G.Y. programme as a whole, and antarctic activities in particular: 16 specialists have written chapters varying in length from 4 to 15 pages within each of 4 main sections devoted to (1) geography and meteorology, (2) geology and structure, (3) upper air physics (including geomagnetism, aurora, ionosphere and cosmic rays) and (4) fauna and flora. A large map of the Antarctic specially prepared by the American Geographical Society is provided.

In their chapter on climatology and meteorology, H. Wexler and M. J. Rubin ask some of the questions which they hope may be at least partially answered as a result of the antarctic I.G.Y. activities: what is the broad-scale wind regime over and around the continent, and does the belt of strong westerlies that ring the continent exclude or substantially reduce the processes of meridional exchange common to middle and high latitudes in the northern hemisphere; how does the antarctic atmosphere work in achieving a near-zero heat balance in particular seasons; if there is ozone in the atmosphere during the winter, how does it get there; does the antarctic tropopause disappear at times of very low surface temperature; and where and when is precipitation desposited ? Answers to these and the other questions briefly discussed in this stimulating chapter would go a long way to elucidate some of the outstanding problems in world meteorology.

A work of this kind compiled with praiseworthy haste in the hope of helping others to an appreciation of the magnitude and importance of the problems that wait to be tackled must have its weaknesses; compare, for example, the different positions given on pages 84 and 94 to the southern zone of maximum auroral frequency. But for the most part shortcomings of this kind serve only to illustrate the urgent need for a concerted effort in collecting reliable observations in the Antarctic and reinforce the value to geophysics of the forthcoming I.G.Y. activities.

The American Geophysical Union and its secretary, Waldo Smith, are to be congratulated on this first geophysical monograph: may the series be long and fruitful.

J.M.S.

Meteorological Office Report for the year ending 31 March 1956, (M.O. 620). London (H.M. Stationery Office). Pp. 59; Tables; 4 Plates. 3s. 6d.

The publication under review yields copious evidence of lively activity within the Meteorological Office. It is not proposed here to try to summarize what is in itself a brief summary of the many activities of the Office. There are, however, some readily recognized high-lights in the year's history. It is good news that there is now a reasonable chance that in the not too distant future the activities now maintained at Victory House in Kingsway, and at Harrow and Dunstable will be collected together in a combined Headquarters. While a general removal and re-assembling may be initially disturbing to many of the staff, it must in the long run result in increased efficiency.

A feature of the report under review is the evidence of the steady growth in the demands from the general public, industry, and the services, for a wide variety both of information and of weather forecasts. The installation of the automatic telephone service (WEA) is an outstanding effort to meet a part, at least, of these demands, while the new edition of *Your weather service* helps the reader to formulate his questions to the meteorologist.

In the section of the report dealing with research it was good to read that the section of the staff at Porton is still interested in problems of turbulence and diffusion, but now with particular emphasis on the greater distances, and with a link with problems of artificial rain-making.

We shall all look forward to the installation within the Meteorological Office of a high-speed electronic computer. This machine will be a valuable research tool, not only for forecasting problems, but also for many other types of meteorological research problems.

The Annual Report under review affords clear evidence that the Meteorological Office is very much alive to the need for continued research. One paragraph alone in the whole Report makes depressing reading, namely, the paragraph on p. 7 which relates to the staff situation, especially in the Scientific Officer and Assistant grades, where recruitment is unhappily much too slow.

D.B.

Einführung in die Optik der Atmosphäre. By Gerhard Dietze. Leipzig (Akademische Verlagsgesellschaft. Geest and Portig K-G), 1957. Pp. XI, 263; 114 Diagrams; 39 Tables; 5 Colour Plates. 29 D.M.

This manual has been prepared by the author from the lectures on atmospheric optics which he gave in the section devoted to meteorology in a general course on the physics of the atmosphere at Leipzig. A knowledge of this subject is important to students of astronomy, geodesy, technical optics, photography and navigation as well as to the meteorologist. The many subjective phenomena of atmospheric optics are of interest to the psychologist and physiologist. This book deals in a general and elementary way with all aspects of the subject. Where possible, the theory is developed using simple mathematics; where more difficult mathematics is essential to an adequate examination of a phenomenon, merely the final formulae are given and their application explained as, for example, in the case of the Mie scattering theory. But in all cases the underlying physical reasoning is given with the utmost thoroughness and clarity and the abundance of illustrative line diagrams is obviously the outcome of much use of the blackboard. The book is excellently produced and will be of great value both to teachers and students. It is churlish perhaps to cavil at the fact that there are only five colour photographs, excellent though these are. This is a student's text-book and the publishers presumably set a limit to its cost. But a manual of atmospheric optics simply cries for copious illustrations by the colour photographs now available for almost all its phenomena, common and rare.

There is as yet no similar book in English. An elementary text on atmospheric optical phenomena combining the formal development of the theory as in this German manual with something of the enthusiastic and human presentation of Minnaert's excellent *Light and colour in the open air* and amply illustrated by colour photographs would surely appeal widely to a multitude of interested laymen as well as to those professionally concerned. May some author and publisher soon provide such a *vade mecum* !

J.P.

Atmospheric electricity. By J. Alan Chalmers. Pergamon Press, 1957. Pp. viii, 327; 65 Fig.; Tables. 63s.

Some eight years ago Dr. Chalmers published a small volume on atmospheric electricity. The present book is recognizably an expansion and a bringing up to date of the first; it possesses, understandably, the same virtues and the same defects as its predecessor.

The reader who, already knowing something of the subject, wishes to make further progress, will find the book of great value. It contains, in itself, a vast amount of information. Furthermore, the bibliography includes over 500 items. Few original papers can have escaped the attention of the author, and any researcher in any facet of atmospheric electricity may be confident that if he studies the references indicated by Dr. Chalmers he will be exceptionally well-informed.

Nevertheless, the comprehensive nature of the volume, besides being its main strength, is also in another respect its major weakness. In presenting the information Dr. Chalmers has been far too kindhearted with his colleagues. Everyone's work is included with equal emphasis, and the succession of brief summaries is at times reminiscent of an appropriate page in *Meteorological Abstracts* ! There is disappointingly little distinction between the important and the trivial and the fundamentals of the subject are correspondingly obscured.

For a book priced at 63s. there are too many typographical mistakes. Errors of fact are less frequent but two are especially surprising. Firstly, it is remarkable to find Dr. Chalmers, so rightly punctilious on the subject of units, perpetuating the common blunder that the unit of columnar resistance is ohm m^{-2} and not, as is correct, ohm m^2. Secondly, in speaking of atmospherics, the section headed ' Whistlers ' is completely confused; the demonstration by Storey of how the signal emitted by a lightning flash can propagate dispersively along a geomagnetic line of force is not even mentioned.

To summarize, no active and experienced worker in atmospheric electricity should be without this well-documented, comprehensive volume; equally, owing to the absence of any discriminative approach, one would hesitate before recommending this book to a fledgling student, whose own critical abilities are, as yet, undeveloped.

E.T.P.

Klima und Bioklima von Wien. II Teil. By F. Steinhauser, O. Eckel and F. Sauberer. Vienna (Austrian Met. Soc.), 1957. Pp. 136; 39 Figs.; 41 Tables. Np.

This is the second part of a comprehensive study of the climate of Vienna, to be completed in three volumes. The first part appeared in 1955 and was reviewed in the *Quarterly Journal*, **82**, 351, pp. 115-6. The presentation of so many aspects of the climate of one city is probably unique and will serve as a model

to town-planners and architects, industrialists and municipal authorities, doctors and meteorologists everywhere.

The first volume of the present work was largely occupied with tables of averages and frequencies of a great diversity of meteorological elements and phenomena.

The second volume now issued contains a simple descriptive text and diagrams, the large diagrams being given as neatly folding double-page sheets stuck in the back cover. The first part of this volume discusses secular variations of temperature, rainfall, snow-cover and wind from the point of view of how far the climatic figures of the past may be expected to be relevant to the future. Of interest to the meteorologists is the revelation that, at least at Vienna, climatic changes over the epoch since 1780 due to meteorological causes have outweighed changes due to growth of the city. Moreover, the changes since the most oceanic climatic phase around 1920 have gone so far that it can be stated, as regards the whole period 1780-1955, that no continued temperature trend is found in any season of the year.

Periodogram analysis was tried, but nothing obviously useful emerged. More interesting is perhaps the discovery that year-to-year temperature variability was notably greater in the period 1775-1873 than between 1874 and 1933, since when the variability has again been much greater with obvious effects on the frequency of extreme years. There may be no parallel to the equable period of our fathers' generation, upon which so much climatology in many parts of the world is exclusively based.

The second section is devoted to the results of special studies, including *inter alia* the incidence of heavy rain, the spread of smoke from the city, the duration of the season for artificial heating of buildings and the amount of radiation received by house walls with different aspects. A final section is devoted to the temperature distribution month by month at thirty different observation points in the city and its environs, including some detailed temperature profiles along a 20 km base-line on cold nights and warm days.

The work is to be completed with a third volume treating the effects of different general meteorological situations on the city and in its streets.

H.H.L.

Klima-Atlas von Rheinland-Pfalz. Bad Kissingen (Deutscher Wetterdienst), 1957. Pp. 37; 9 Figs.; 75 charts. 32 DM.

The 4th in the series of climatic atlases of the German Länder has now appeared and deals with Rheinland-Pfalz. The same series of many-coloured charts as in the earlier numbers of this series of atlases is provided, while the textual comment and tables of extreme values form a comprehensive survey of the climatic conditions of the west and south-west regions of the country.

R.S.R.

Sir Richard Gregory. By W. H. G. Armytage. London (Macmillan), 1956. 8 vo. Pp. vii, 241; 1 plate. 21s.

In this book the Professor of Education at Sheffield University sets out to tell the life story of a very remarkable man. Richard Arman Gregory, son of a boot-repairer to Clifton College, started his career as a Bristol newsagent's delivery boy at 4s. 6d. per week; eventually he became something of a national figure, having edited *Nature* for 20 years, achieved a baronetcy, and received the high honour of Fellowship in the Royal Society under the rule which provides for the unsolicited admission of ' persons who either have rendered conspicuous service to the cause of science or are such that their election would be of signal benefit to the Society.' This biography, states Prof. Armytage, emerged from posthumous sorting and sifting of Sir Richard's papers and correspondence. By free confession to ' manifest inadequacies ' in his record, thus derived, the author largely disarms critisicm. Nevertheless, for the benefit of future editions, some of the book's defects should be pointed out. Not least of them is the lack of any reference to Gregory's presidency of the Royal Meteorological Society in 1928-30. As was mentioned in the obituary notice which appeared in Vol. 79 (p. 183) of this *Journal*, his second presidential address probably won an extent of publicity unique in the Society's history.

It must also be complained that Prof. Armytage's volume is overloaded with quotations from letters which passed between Gregory and his lifelong friend H. G. Wells. To Gregory's first wife, we learn, Wells was a social pariah; in that opinion she was by no means alone. To Gregory himself Wells seemed more of a paragon than a pariah – ' a modern St. Paul,' ' the greatest international scientific educator of modern times.' Gregory, says Prof. Armytage, was the only person with whom Wells never quarrelled. The two men corresponded amicably and often at great length, over half a century. But this does not justify the author in so constituting his book that Wells requires twice as much space in the index as Gregory himself. The most ardent Wellsian would probably allow that he gets an undue share of attention. Others may find the constant references to his ways, aims and creed rather irksome.

Though thus to some degree unbalanced and incomplete, the biography may be recommended as very readable and full of interest. Among the little-known details of Sir Richard's activities here revealed is that after his second marriage, when in his seventh decade, he took dancing lessons and became a member of Ciro's. Slips and misprints in the book appear to be few; they include ' Aurigia ' for ' Auriga ' on p. 22 and ' 31 September 1946 ' on p. 206.

E.L.H.

Obituary Notices

Prof. Dr. HEINRICH FICKER

Professor Dr. Dr. *h. c.* Heinrich Ficker, who died at Vienna on 29 April 1957, at the age of 75, was one of the great pioneers who created modern synoptic meteorology. He was born in Munich in 1881. After having studied meteorology at the University of Innsbruck under the guidance of Trabert he came to the Central Meteorological Institute at Vienna, where he met J. V. Hann, Pernter, Exner and Margules. In 1911 he received a professorship at Graz. From 1923 to 1937 he occupied a chair at the University of Berlin and was director of the Prussian Meteorological Service. In 1937 he returned to Austria as director of the Austrian Meteorological Service and was professor at the University of Vienna until his retirement in 1953.

Dr. Ficker produced a great number of outstanding contributions to synoptic meteorology and climatology in more than 140 publications. Already in his doctorate thesis ' Innsbrucker Foehnstudien ' he presented a thorough study on foehn winds over the Alps and was able to find theoretical explanations for the observations during the different stages of this phenomenon. He also gave a satisfactory synoptic theory for the pressure distribution necessary to set in motion foehn winds. His work will always be fundamental to future research on the same lines. In 1910 appeared his paper on ' Ausbreitung kalter Luft in Russland,' wherein he gave a complete description of the life history of cold-air outbreaks over Russia. In this paper he introduced for the first time the idea of a ' Polar-front,' some years prior to the famous Norwegian theory. During the years 1920-1923 he made investigations on ' Zusammensetzung von Depressionen ' and was able to show that the development of depressions can only be understood as the interaction of low- and high-pressure waves proceeding in the same direction but usually with different speeds. It is well known that this idea has proved to be very successful in later research. The rôle of upper-air currents in the development of surface depressions is now – with some modifications of Ficker's concept – generally accepted. Once again Ficker made an important contribution to practical weather forecasting when he introduced the steering principle into synoptic meteorology in his investigation of the big storm over Germany on 4 July 1928. Among his other numerous studies his research on trade winds and thunderstorms should be mentioned especially. Dr. Ficker was also an outstanding teacher, and a large number of former students will remember his inspiring personality. His overflowing sense of humour and his kindliness gave him a magnetic appeal to everyone who had the pleasure of meeting him. Dr. Ficker held membership in many scientific societies. He was elected an Honorary Member of the Royal Meteorological Society in 1950.

H.R.

Captain G. BYERS

We have to report the death, in March 1957, of Captain Gerald Byers, a Life Fellow of the Society, elected on 19 June 1912. He joined the Mercantile Marine in 1894 and was for a long period with the China Navigation Company, from which he retired in 1927. During the war of 1914-18 he saw active service on one of the Company's ships in the Mediterranean. After the renewal of hostilities in 1939 he broke his retirement to undertake coastal observation work in Devon. Captain Byers had been keenly interested in weather study for many years and gave a good deal of time to the meteorological records which he kept in the garden of his home at Dartmouth.

E.L.H.

Corrigenda to July 1957 issue

G. B. Tucker—

p. 298, Table 4, column 4, 2nd row—
for + 0·4, *read* − 0·4

p. 299, line 8—
for meridional, *read* zonal

Eq. 1 should read :

$$\frac{du}{dt} = lv - \frac{1}{\rho}\frac{\partial p}{\partial x} + \frac{1}{\rho}\frac{\partial \tau_{xz}}{\partial z} + \frac{1}{\rho}\frac{\partial \tau_{xy}}{\partial y}$$

p. 300, line 1—
for $\frac{\partial v}{\partial t}$, *read* $\frac{\partial u}{\partial t}$

line 2—
for $\frac{dv}{dt}$, *read* $\frac{du}{dt}$

p. 300, Fig. 3—
The sign of div $\bar{v}$ is reversed,
i.e., negative values of div $\bar{v}$ should be positive and vice versa.

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The essay, which must be in English, may include material which the candidate has already published. Essays may be submitted jointly in the names of more than one author.

Candidates are required to submit three copies of their essay, which must be printed or typewritten. In the latter case, only one side of the paper should be used, double spacing should be employed and two of the copies may be good carbon copies. Unless the essay is printed, any mathematics should be written in carefully by hand. Although the scale of the essay will be considerably greater than that of an ordinary paper, candidates are recommended to consult, for general advice, the pamphlet, 'Preparation of papers for the Quarterly Journal.' This may be obtained, free of charge, on application to the Assistant Secretary.

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All entries for the competition should be addressed to the Assistant Secretary, Royal Meteorological Society, 49 Cromwell Road, London, S.W.7 and marked clearly 'Napier Shaw Prize.'

Essays for the competition must be received at the Society's Offices *not later* than 1 January 1959.

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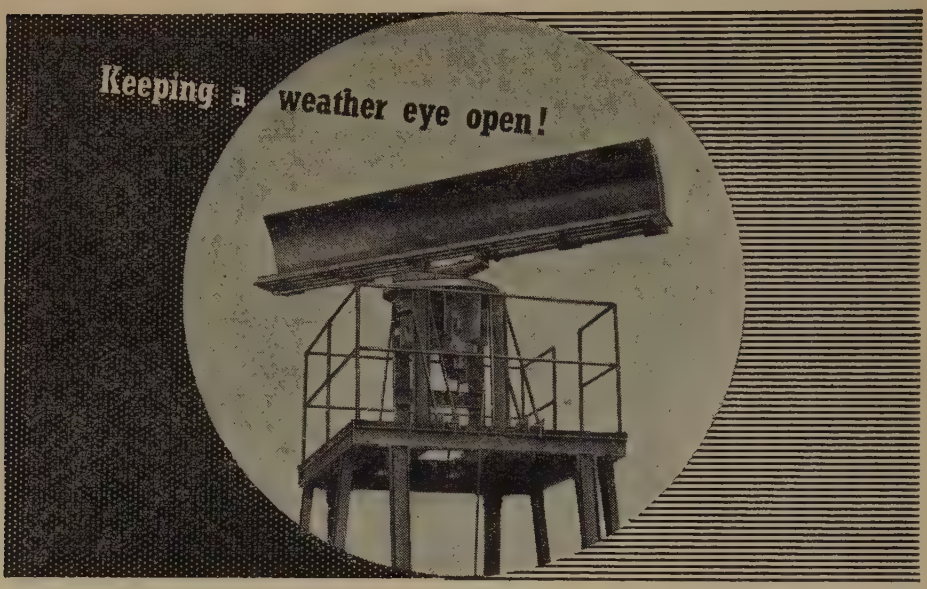

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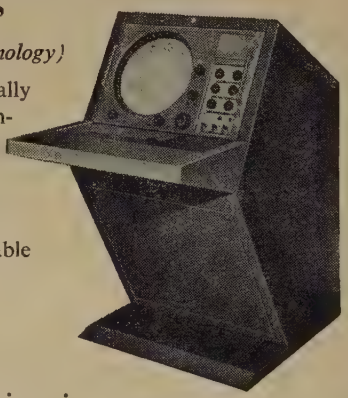

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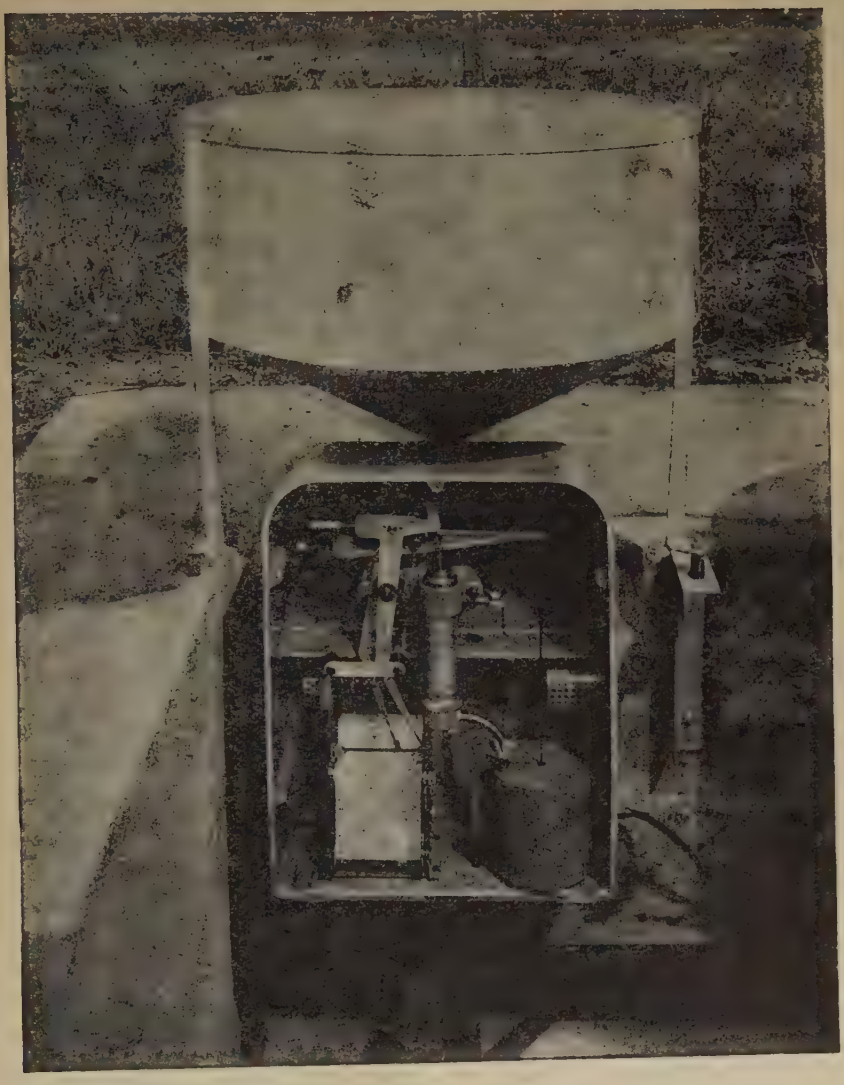

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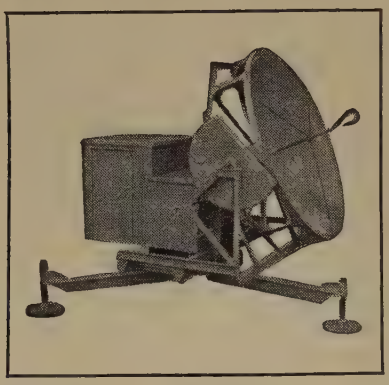

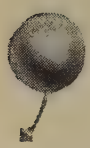

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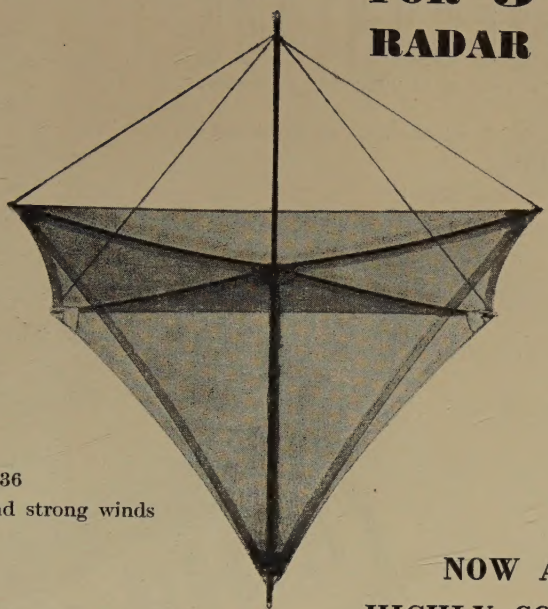

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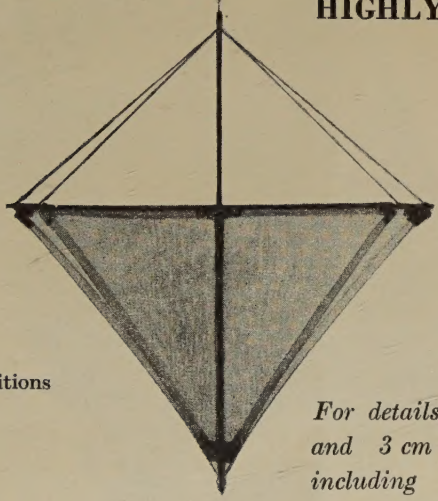

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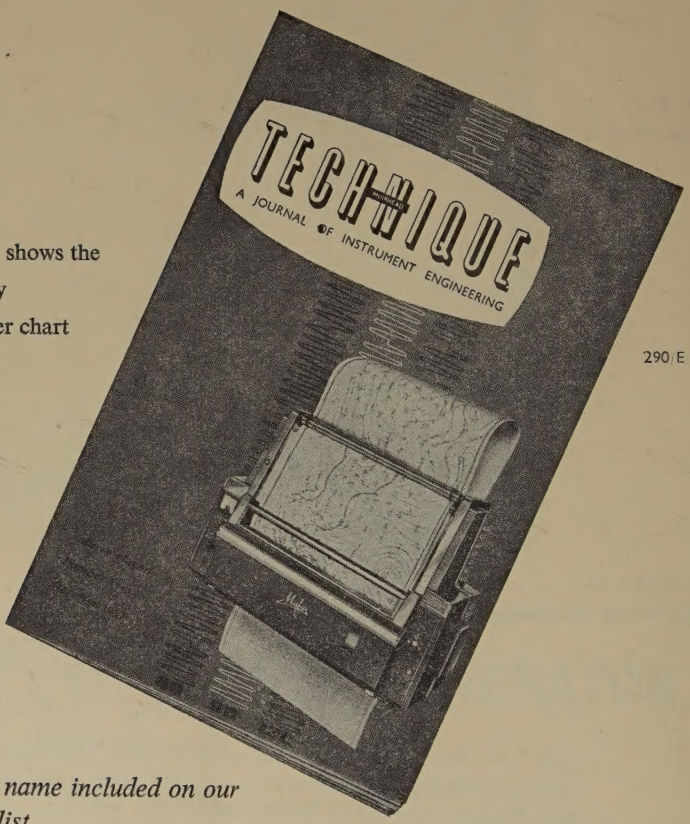

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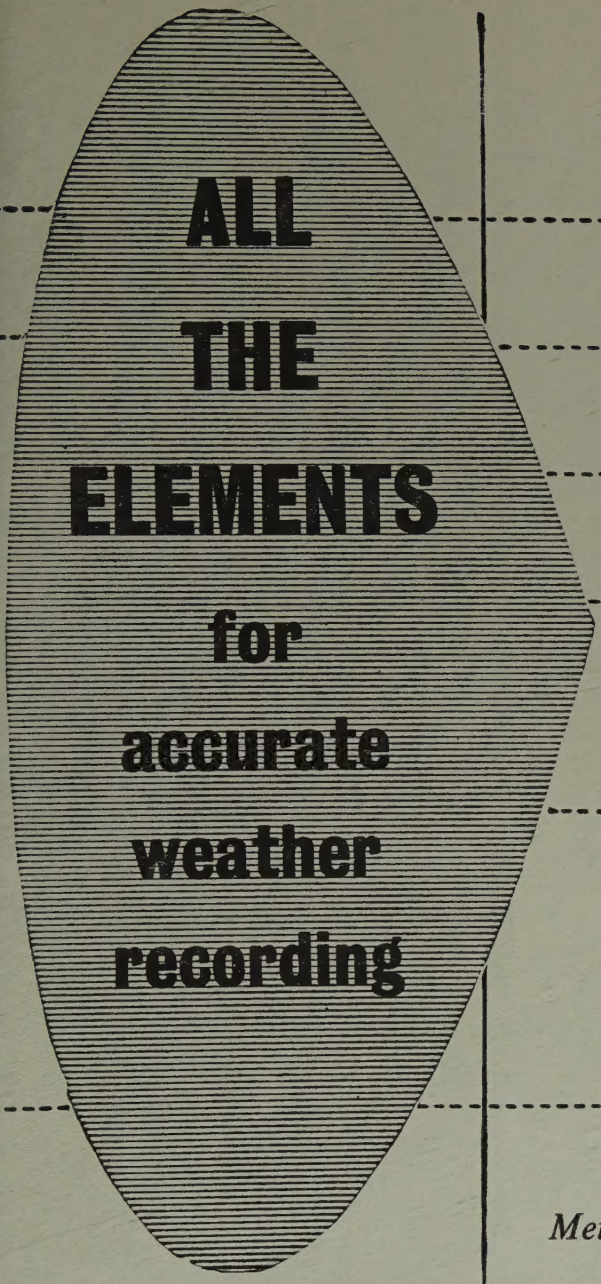

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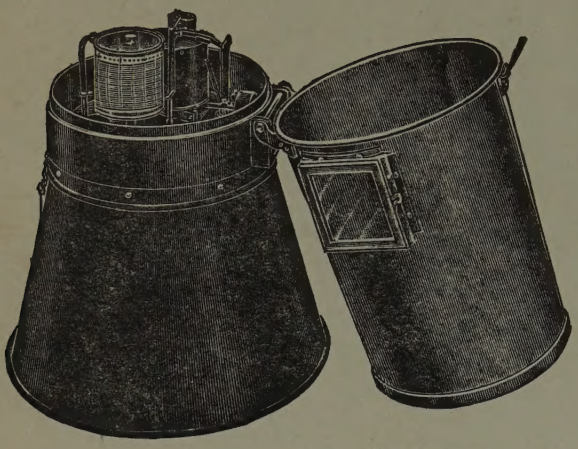

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